

PHYSICAL PROPERTIES OF HOST AND COVER ROCKS

This section summarizes the physical properties of host or cover rocks important for modeling of geophysical responses, or evaluation of geophysical data. The properties considered are density, porosity, magnetic susceptibility and remanence, seismic velocity, resistivity, IP effect, electrokinetic coupling coefficients, thermal conductivity, inertia and heat sources, and radioelement content. This is not a critical summary that tries to evaluate the adequacy of published data. It is simply a summary for which critical evaluation is left to the user. However, the limitations of many published catalogues of physical properties need to be pointed out, because most property measurements are of intrinsic properties derived from measurements on laboratory samples. Properties measured on hand-sized laboratory samples may or may not be representative of those properties in-situ. Particular care needs to be taken if laboratory measurements of density, seismic velocity or resistivity are extended to represent in-situ bulk properties because of macro-scale fractures that may be present in the earth and the amount of interstitial water.

In choosing host or cover rock properties from lists such as presented here the effects of alteration processes on those properties need to be considered. Many of the values represent measurements on relatively fresh samples of rock. Processes such as weathering, diagenesis, metasomatism and hydrothermal alteration can significantly affect all of the physical properties, in some cases causing an increase in value and in others a decrease. Processes attendant to mineralization often work to the advantage of the explorationist by providing a larger and more easily identified indirect geophysical target. Alteration processes generally increase the range of possible property values, add to the geologic noise and may make interpretation more difficult.

Density

A number of factors affect the in-situ density of earth materials, including porosity, water content, depth of burial, age, crystallinity and chemistry. Awareness of all these factors is important in evaluating the applicability of the gravity method to a particular exploration problem, especially in regard to the magnitude of "geologic noise" to be expected. Sediments, both unconsolidated and consolidated, may vary considerably in density depending on the degree of saturation. Density contrasts can exceed 0.5 gm/cm³ between wet and dry sand or gravel. Figure 5 shows ranges of densities, wet and dry, and porosities for sandstones, shales, limestones, dolomites, and unconsolidated material of various types from three literature sources (Telford and others, 1976; Jakosky, 1950; and Fedynskiy, 1967). Most apparent from figure 5 is the wide range of densities exhibited by these sedimentary units. This presents obvious problems to the modeler in knowing what densities to choose where sediments are present. Note also the high densities shown by some shales.

Based on literature values, the density variation for individual igneous rock types is not nearly as large, as shown for sediments. Figure 6 shows the ranges of bulk density from four literature sources (Daly and others, 1966, Telford and others, 1976; Johnson and Olhoeft, 1984; and Mironov, 1972), for 13 igneous rock types. Because of the low porosity of most igneous rocks, there are only minor differences between wet and dry densities. Many references do not indicate whether wet or dry densities were measured, and in some cases original sources are not indicated. Similarities in values between the references shown and other sources suggest that many of the data date back to Daly's 1933 work and Reich's 1914 work (Heiland, 1940).

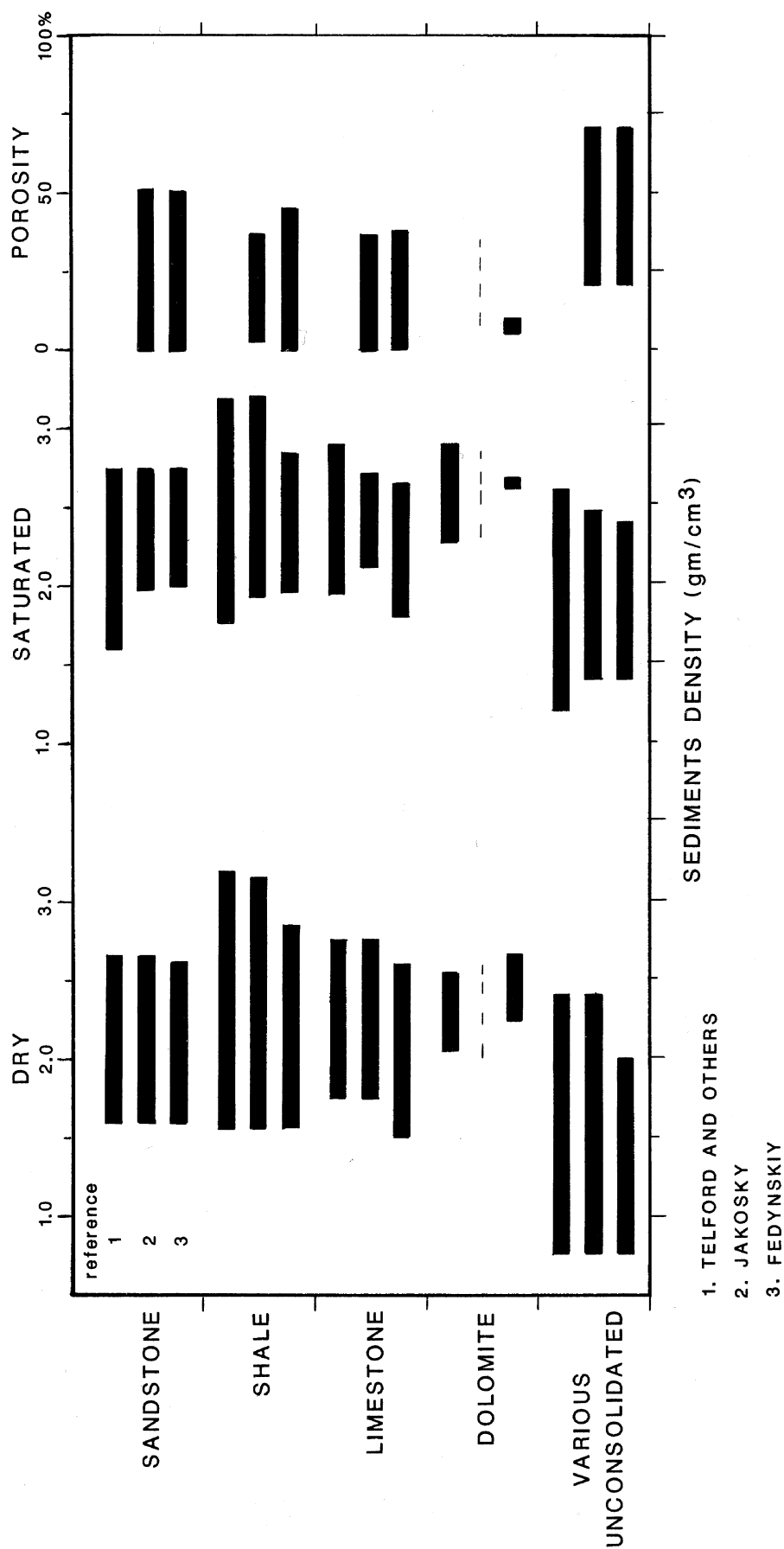


Figure 5. Diagram showing ranges of wet and dry bulk densities and porosity for various sedimentary rocks. Reference sources are 1. Telford and others, 1976, 2. Jakosky, 1950, and 3. Fedynskiy, 1967.

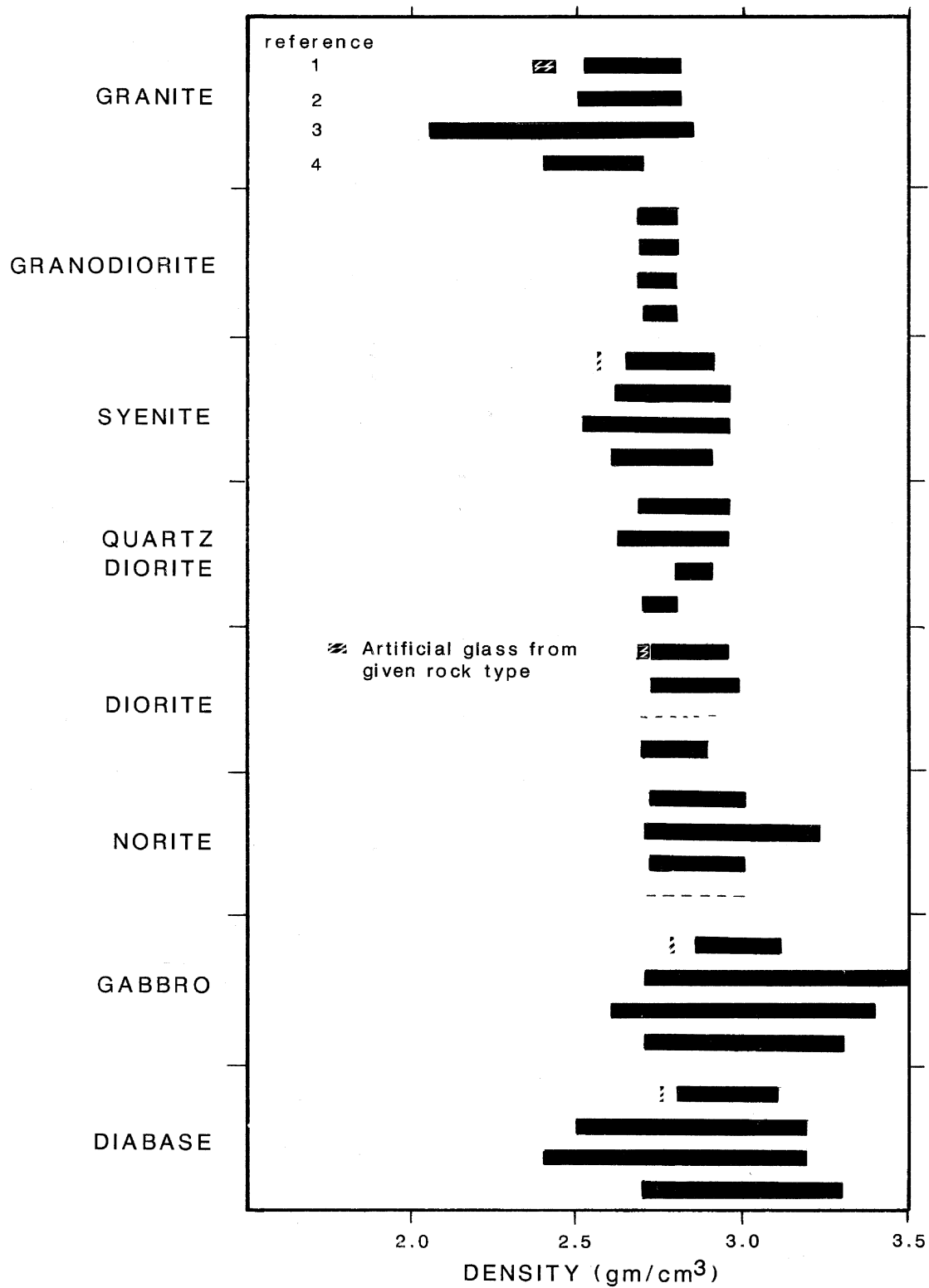


Figure 6. Diagram showing ranges of bulk densities for 13 different igneous rocks. Reference sources are 1. Daly and others, 1966, 2. Telford and others, 1976, 3. Johnson and Olhoeft, 1984, and 4. Minonov, 1972.

The values shown in figure 6 appear to represent relatively fresh samples. Alteration from weathering and some but not all hydrothermal processes would reduce the density. The wide range shown for granite by Johnson and Olhoeft (1984) in figure 6 probably results from inclusion of altered samples among the 334 granites measured.

In figure 6, density values are also given for artificial and natural glasses from Daly and others (1966). Five artificial glasses derived from granites, syenites, diorites, gabbro and diabase are shown next to their counterparts. The glasses are all significantly less dense than their igneous equivalent. Extrusive igneous rocks are typically less dense than their intrusive equivalent in part because of their glass content, and often because of increased porosity.

Figure 7 presents the ranges of densities for 9 metamorphic rocks as reported from the same sources used for igneous rocks. Note the relatively low range reported for slate. As with the igneous rocks many reference sources do not indicate whether the measurements are of wet or dry samples. In most cases these metamorphic rocks also would be of low porosity, if fresh, causing little change between the wet and dry state. Johnson (1983) has tabulated dry bulk density and porosities for 182 different rock samples. Table 2 provides a summary for several types of metamorphic rocks from his work. The porosities do not exceed 3.49% on his samples. Further details on rock densities can be found in the references cited.

| Table 2. Dry bulk density and water available porosity of selected metamorphic rocks from Johnson (1983). | | | |
|--|----------------------|-----------------------|-----------------------|
| Rock type | Density range | Porosity range | No. of samples |
| Quartzite | 2.645-2.733 | 0.11-0.36 | 3 |
| Hornfels | 2.688-2.709 | 0.07-0.29 | 2 |
| Schist | 2.634-3.018 | 0.29-3.49 | 9 |
| Marble | 2.648-2.991 | 0.07-0.7 | 9 |
| Slate | 2.762-2.79 | 0.25-0.43 | 4 |
| Gneiss | 2.617-3.137 | 0.22-1.73 | 7 |
| Eclogite | 3.251-3.359 | 0.09-0.36 | 2 |

Porosity

Porosity is a property that is not measured directly by one of the geophysical techniques, but one that can dramatically affect in-situ density, resistivity and P-wave seismic velocities. It is because of the effect on these other properties that porosity is important in interpretation and modeling of ore deposits. Literature on the geophysical attributes of ore deposits infrequently provides measured porosity data, and this is reflected in the presented models by the few quantitative values for this parameter.

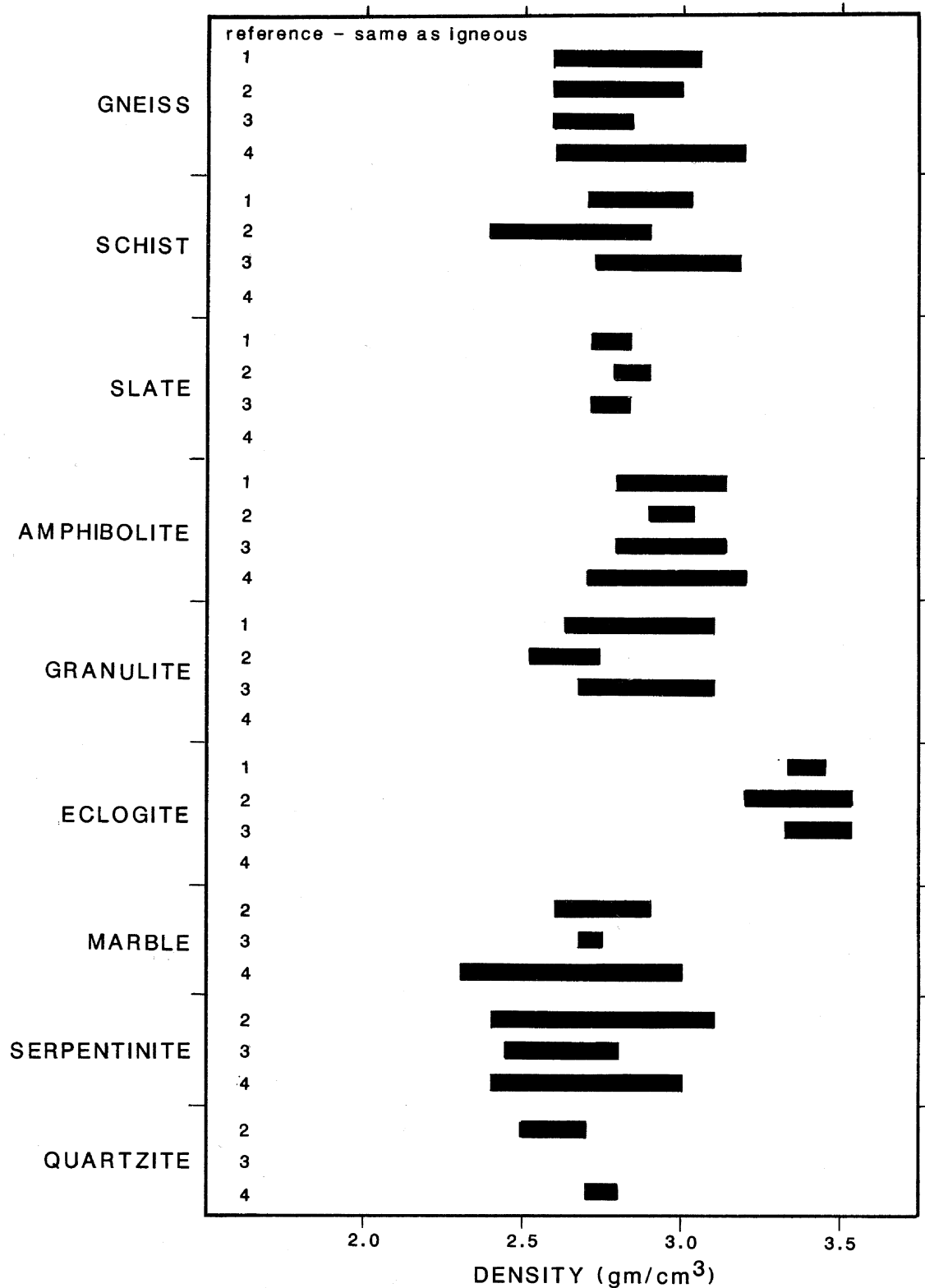


Figure 7. Diagram showing ranges of bulk densities for 9 different metamorphic rock types. References sources are 1. Daly and others, 1966, 2. Telford and others, 1976, 3. Johnson and Olhoeft, 1984, and 4. Mironov, 1972.

There are two types of porosity of primary concern for their effect on rock properties, intergranular, and joint porosity. With the exception of chemical sediments, sedimentary rocks typically have high intergranular porosity. Joint or fracture porosity in all typical rock types is small in magnitude but can be important in its effect on rock properties. Table 3 compares ranges of intergranular and joint porosity for several groups of rocks as given by Keller and Frischknecht (1966).

Borehole logging methods, such as neutron, resistivity and seismic velocity, provide estimates of porosity, but these are primarily used in petroleum work. Direct knowledge of porosity may become increasingly important as the process of in-situ leaching of metallic ores is increasingly utilized.

Porosity is important to the exploration geophysicist principally for the part it plays in affecting density, electrical resistivity, and seismic velocity. Changing from water saturated to dry conditions may decrease densities by 0.5 gm/cm³, may increase resistivities several orders of

| Table 3. Ranges of intergranular and joint porosity for several rock types. Data from Keller and Frischknecht (1966). | | |
|--|---------------------------------|-------------------------|
| Rock type | Intergranular porosity % | Joint porosity % |
| Precambrian igneous and higher grade metamorphic | 0-2 | 0-2 |
| Paleozoic and younger igneous | 0-10 | 0-2 |
| Precambrian sediments and low-grade metamorphic | 1-8 | 0-2 |
| Paleozoic sandstone and shale | 5-30 | 0-1 |
| Paleozoic limestone | 2-10 | 0-2 |
| Paleozoic clastic volcanic | 5-30 | 0-2 |
| Younger sandstone and shale | 10-40 | 0 |
| Younger limestone | 4-20 | 0-2 |
| Younger clastic volcanic | 10-60 | 0 |

magnitude, and may decrease seismic P velocities by a factor of 2 or more. Even low porosity-rocks may show significant changes in resistivity and seismic velocities between wet and dry conditions. Christensen (1982) lists a granite with 1.1% porosity whose P-wave velocity decreased by 40% from wet to dry conditions. The effects of water saturation on these other physical properties is covered in the individual sections dealing with these properties.

Porosity values for different rock types can be found scattered throughout the physical property literature often in conjunction with density, resistivity, or seismic velocity tabulations. Daly and others (1966) provide porosity ranges and averages for a variety of sediments and Christensen (1982) gives density, porosity, and seismic velocity for a variety of marine sediments and igneous rocks. The data of Daly and others (1966) are

summarized in figure 8. These data can be compared with porosity values for sediments given in figure 5. Intrusive and high grade metamorphic rocks typically have porosities of a few percent or less when fresh and unfractured. Fracturing and alteration, however, can increase in-situ porosity significantly, especially where clay minerals are formed. Silicification, on the other hand, often will decrease porosity. As such processes are common during mineralization from hydrothermal systems, their effects on ore and adjacent host rock need to be considered for modeling. Porosities of extrusive rocks can vary widely, but compilations of representative values are not given in the standard references. Johnson (1983) provides some values for these types of rock, ranging from 0% for obsidian to 62.4% for pumice.

It should be remembered that high porosity does not necessarily imply a high permeability for a rock unit. Clays are excellent examples of a lithologic type with high porosity, but very low permeability. There are no generalities that can be made regarding the relationship between porosity and permeability, other than that for rocks of a given porosity the permeability will, in general, decrease with decreasing grain size. Johnson (1983) gives measurements of both porosities and permeabilities made on the same samples of a wide variety of rock types. His data for water available porosity and permeability have been plotted in figures 9a for igneous rocks, and 9b for sedimentary rocks. The wide scatter shown is indicative of the lack of correlation between these two properties. The graphs, figures 9a and 9b, are useful for showing ranges of these two properties as measured on hand-sized specimens.

Magnetic Susceptibility and Remanence

The magnetic properties of rocks depend on the quantity, composition, grain size, and physico-chemical history of magnetic minerals normally present as minor constituents of a rock unit.

Iron ores are the major exception where magnetic minerals can form the bulk of the rock and thus provide a strong target for geophysical exploration. Various authors have provided formulae relating the magnetite content of iron ores and rocks to susceptibility. Werner (Hansen, 1966) derived two expressions, one for Swedish magnetite ore and the other for hematite ore and other rocks. These are, for magnetite ore;

$$k = \frac{Kv}{1 + cKv} \quad (1)$$

where

$$c = \frac{4\pi}{3} \frac{1 - v^{1/6}}{v} \quad (2)$$

k = ore susceptibility
 K = susceptibility of magnetite present
 v = volume fraction of magnetite present

and for hematite ore, igneous or sedimentary rocks;

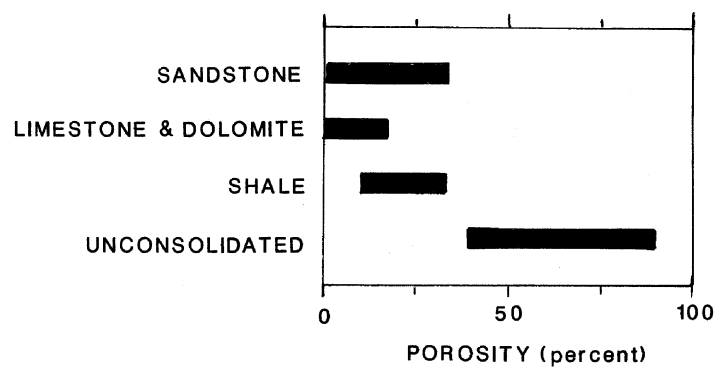
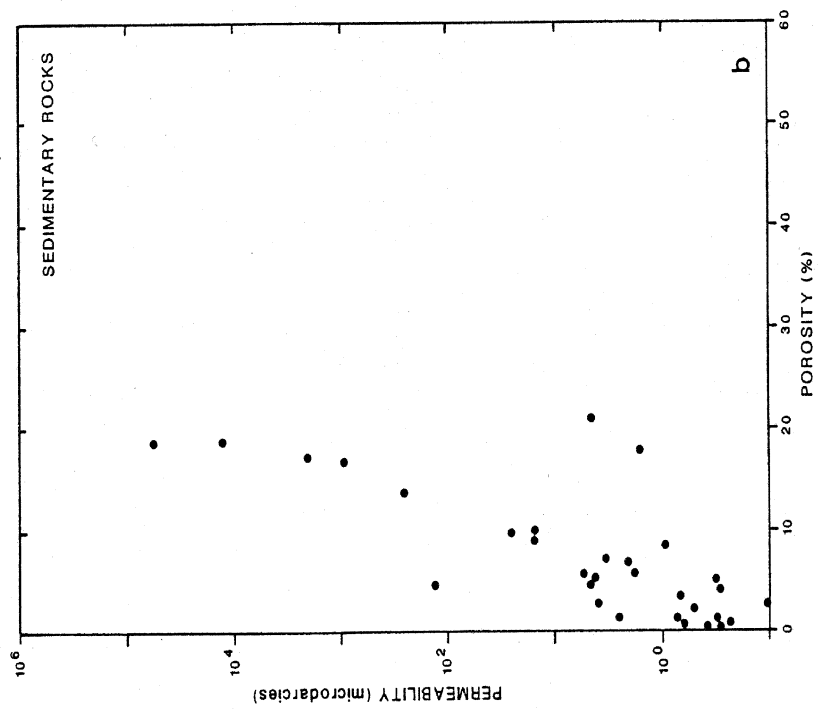
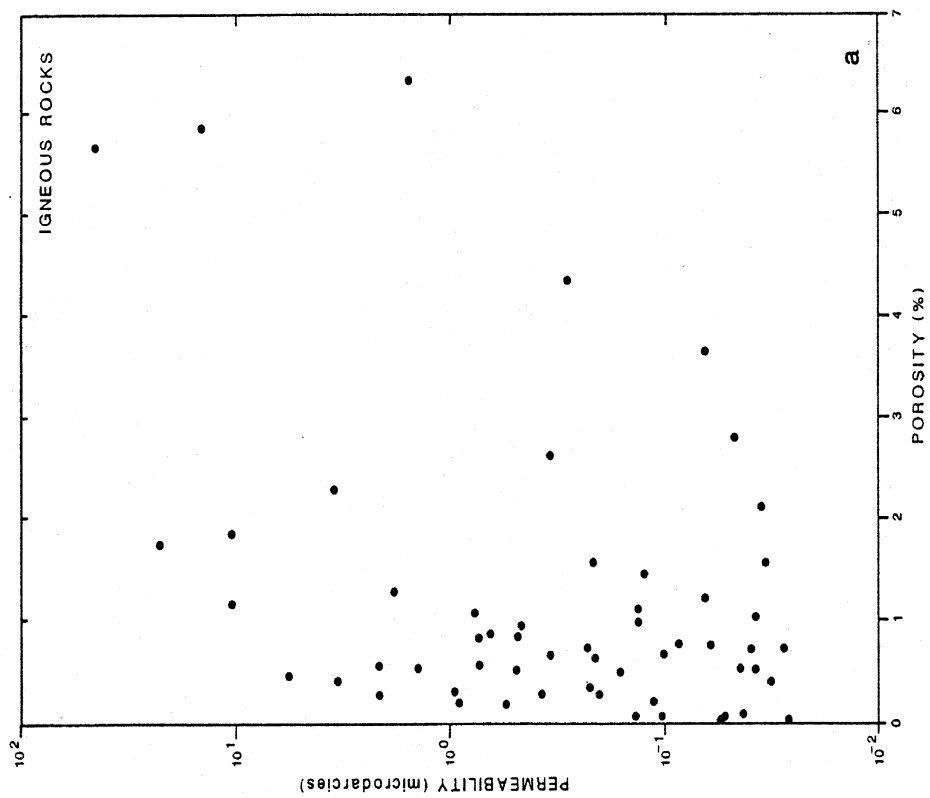


Figure 8. Diagram showing ranges of porosities for sedimentary rocks from Daly and others (1966).



Figures 9a and b. Measured porosity and permeability on a suit of a, igneous rocks and b, sedimentary rocks adopted from Johnson (1983).

$$k = \frac{Kv}{1 + ckv} + 120(S - S_1) \times 10^{-6} \quad (3)$$

where

S = specific gravity of the non-magnetite part of rock
 S₁ = specific gravity of the quartz, feldspar, or limestone fraction of the rock.

Grant and West (1965) summarize results from Mooney and Bleifuss (1953) on Precambrian rocks from Minnesota, and from Balsley and Buddington (1958) on Adirondack rocks. Bath (1962) and Jahren (1963) give results on Minnesota iron formation ores, and Dukhovaki (Klichnikov and Benevolenskiy (1970) on granites from the Akchatausk complex of the USSR. The relationships given by these authors are given below and shown in figure 10.

- | | | |
|--|--|-----|
| 1. Mooney and Bleifuss (1953) | $k = 2.89 \times 10^{-3} V^{1.01}$ | (4) |
| 2. Balsley and Buddington (1958) | $k = 2.6 \times 10^{-3} V^{1.33}$ | (5) |
| 3. Bath (1962) or Jahren (1963) | $k = 1.16 \times 10^{-3} V^{1.39}$ | (6) |
| 4. Klichnikov and Benevolenskiy (1970) | $k = 2.38 \times 10^{-3} V, V < 9.6\%$ | (7) |

These relationships are much simpler than those given by Werner and should provide adequate approximate estimates of susceptibility where magnetite content, V, of a rock is known or can be estimated.

A well known rule-of-thumb is that the more mafic a rock is, the greater is its associated susceptibility (Carmichael, 1982; Grant and West, 1965; Slichter, 1942). This has been expressed as (Carmichael, 1982) basic extrusive > basic intrusive > acid igneous > sedimentary. This relates to the greater magnetic-mineral content of mafic rocks. Where this content is relatively coarse-grained the mafic rock will have a large magnetic susceptibility and if fine-grained it will have a relatively large remanent magnetization. This relationship is documented in various handbooks and texts to which the reader can refer for details. Figure 11 shows the ranges of susceptibilities given by three of these references for igneous rocks. Average values for each group are indicated by a mark on the range bar, and show the increased susceptibility of mafic rocks. The wide range of susceptibilities for a given rock type is also apparent.

Figure 11 shows a wide range of susceptibilities for granites, but does not distinguish between the various varieties of granites. Chapel and White (1974) have defined I-type, for igneous, and S-type for sedimentary, granites based on their chemistry and inferred source rock. Ishihara (1977) has defined ilmenite series and magnetite series granites based on the opaque mineral content (ilmenite-series granites by definition contain less than 0.1% magnetite). These classifications are important in exploration because I-type or magnetite-series granites are generally associated with gold and base-metals while the S-type or ilmenite-series granites relate to tin and tungsten mineralization. A compilation of the variation of susceptibility and other physical properties of these varieties of granites would be desirable, but none is known to the authors. Both the S-type and ilmenite-series granites have low susceptibilities. Based on the definition of the ilmenite series granites, figure 10 can be used to estimate the maximum susceptibility.

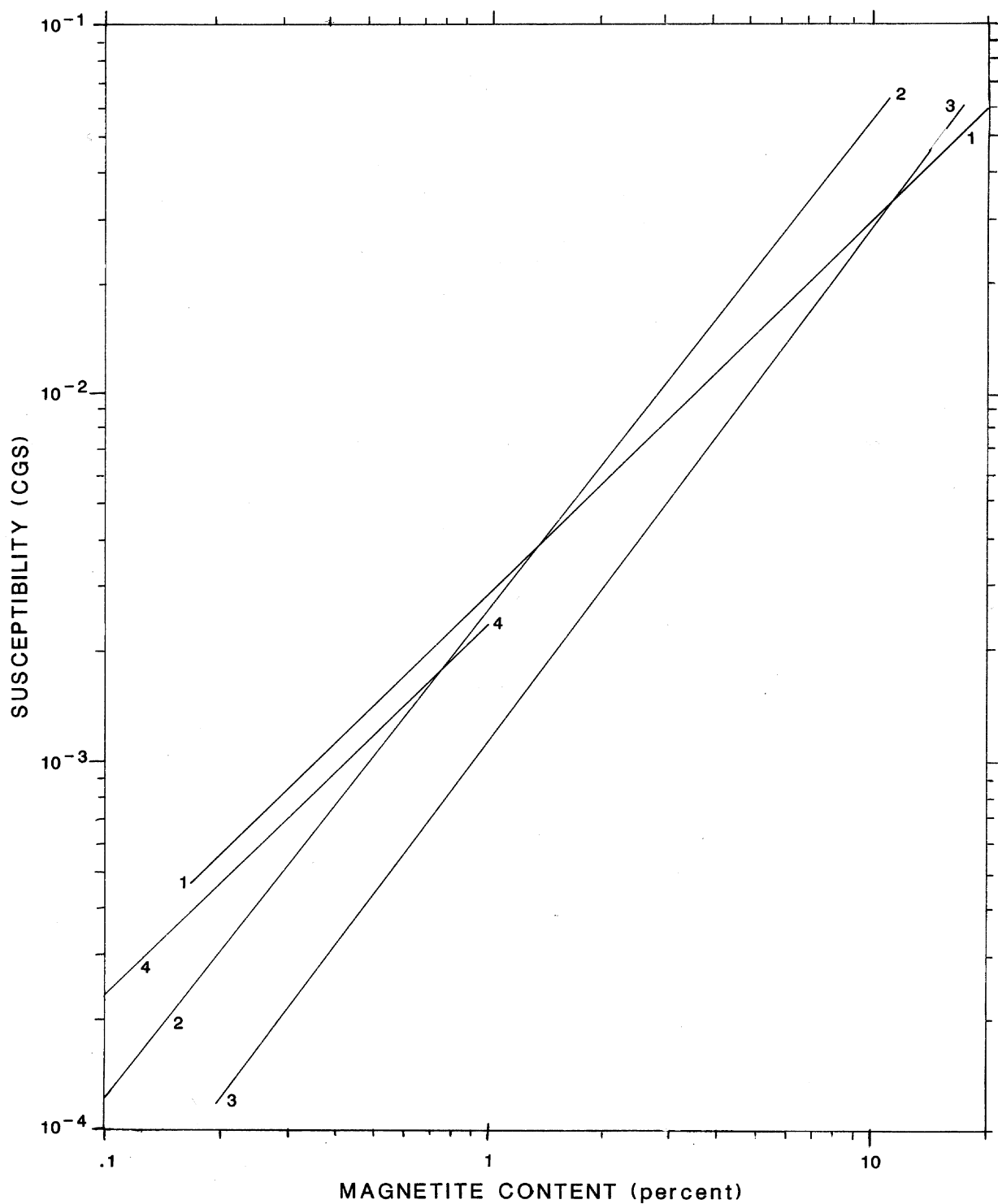


Figure 10. Curves showing empirically derived relationship between magnetic susceptibility and magnetite content from 1. Mooney and Bleifuss (1953), 2. Balsley and Buddington (1958), 3. Bath (1962) or Jahren (1963), and 4. Klichnikov and Benevolenskiy (1970).

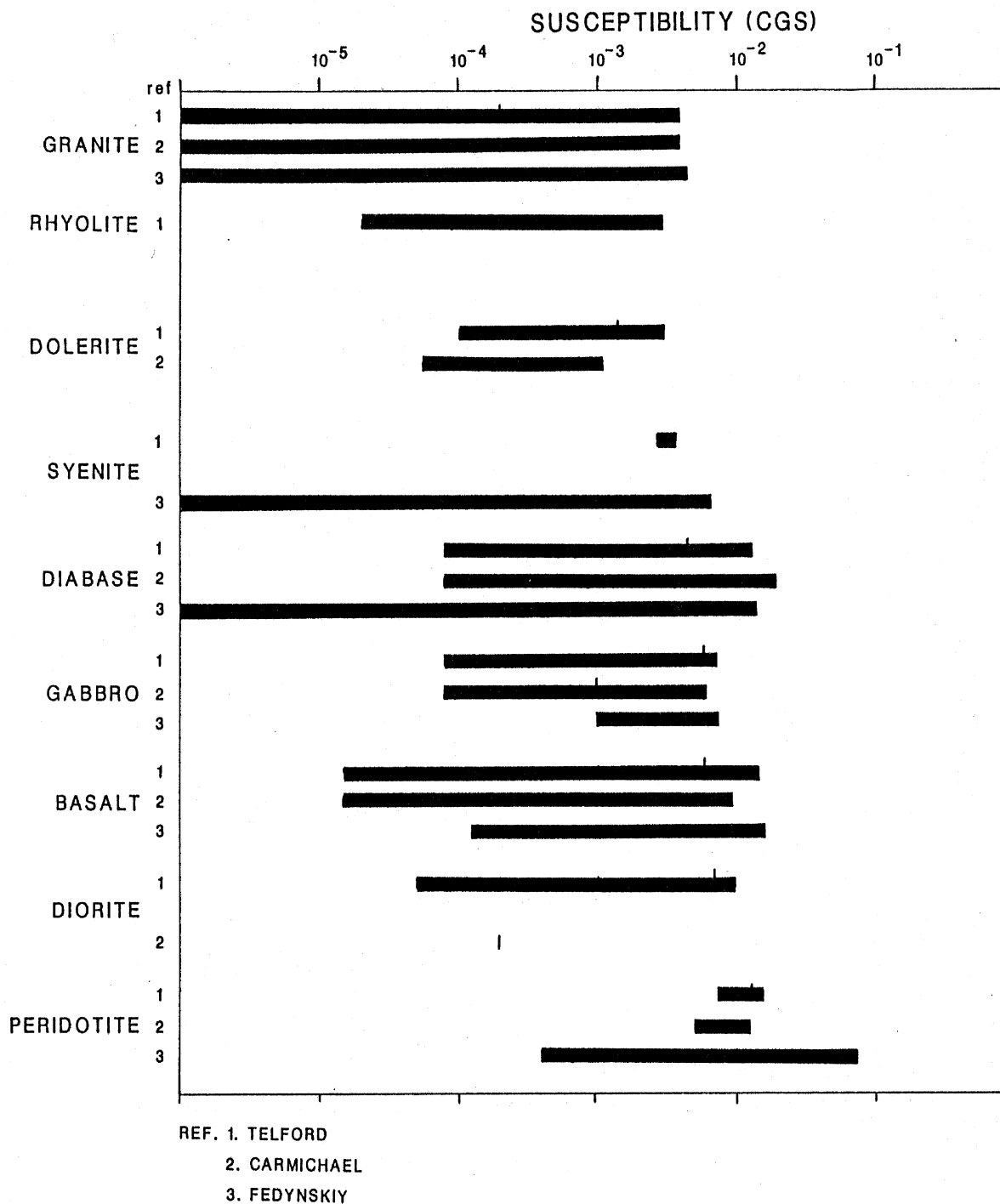


Figure 11. Diagram showing ranges of susceptibilities for various igneous rocks. Reference sources are 1. Telford and others, 1976, 2. Carmichael, 1982, and 3. Fedynskiy, 1967. A bar on the data of Telford and others (1976) indicates the average value.

Figure 12 illustrates the ranges of susceptibilities of sedimentary units taken from four sources identified in the figure. This figure shows the generally lower susceptibilities associated with these rock types. In this figure soils are seen to have a wide range of susceptibilities reaching as high as 10^3 cgs/cm³. Near surface soil zones have been found in many cases to have higher values than lower zones, this is attributed to the formation of maghemite (see Grant and West, 1965 for a summary). Such soils could be a significant source of "geologic" noise in ground surveys.

Besides magnetite ore, several other ores have moderate to high susceptibilities that may permit direct detection by magnetic methods. Figure 13 shows the ranges of susceptibilities reported for eight ores by Carmichael (1982) and Parasnis (1966).

Remanent magnetization, or the related Koénigsberger ratio Q , which is the ratio of remanent to induced magnetization, is not often determined as part of an exploration program for minerals. Thus relatively few data of this important property are available for ore deposits. This is reflected in the compilations of deposit models by the lack of quantitative data. This property is important for interpretation of magnetic data because some anomalies are due in greatest part to remanence. Extrusive volcanic rocks typically show strong remanence or high Q values, but other rocks often show it as well. Hawes (1952) describes studies of the Spavinaw granite in Oklahoma that exhibits a large aeromagnetic anomaly for a granite, and which shows Q values above 100.

Studies of remanence, and values for various rock types have principally been produced from paleomagnetic studies. Such studies have potential for mineral deposit exploration and investigation but are not commonly used in exploration. Hood (1961) has presented a study of the Sudbury basin where he showed that the norite in the mafic complex can be identified by its consistently high Q . Another study by Gross and Strangway (1966) showed how remanent magnetization studies can help address the genesis of iron formation ore.

Details of the various types of remanence, the stability of each, and effects important to the explorationist are beyond the scope of this study. For those wishing to look further into the subject, Grant and West (1965) provide a good overview.

Table 4 lists Q values for selected rock types from Carmichael, 1982 and Kuz'micheva and Diomidova (1968).

Seismic velocity

Investigations of seismic velocities as a function of rock type have been most extensively studied for sedimentary rocks because of the importance of seismic methods in oil and gas exploration. The geophysical literature is replete with details on velocity distribution as functions of porosity, depth of burial, and fracturing in the sedimentary section. There are few in-situ studies of the velocities of ores and host rock environments that are important in minerals investigations. However, laboratory measurements and empirical relationships developed from them do permit reasonable estimates to be made of the velocities to be expected in the minerals environment.

This summary will give ranges of values for seismic P-wave velocities only. Although S-wave techniques are finding increasing applications, they do not appear to have had any significant application yet in minerals

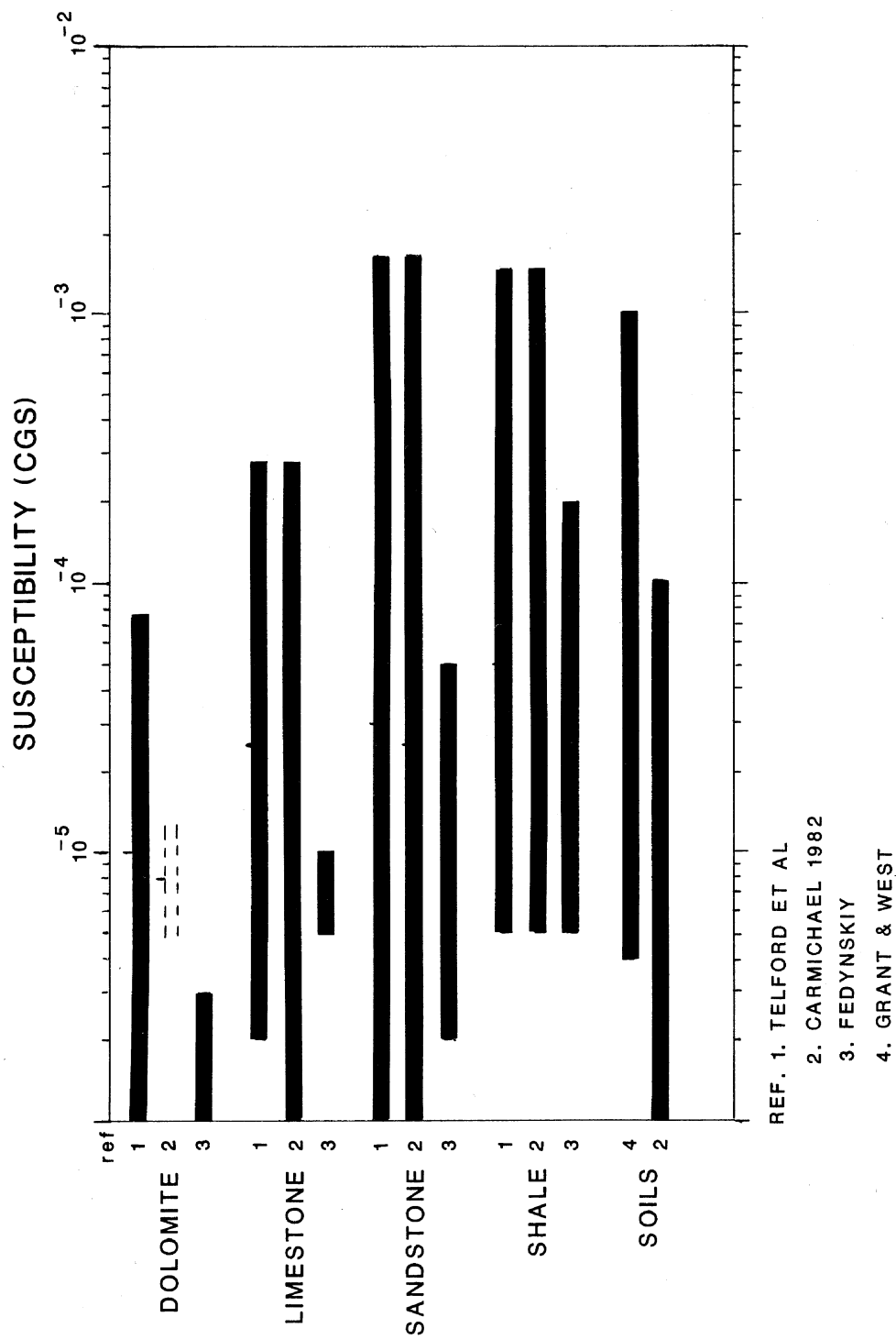


Figure 12. Diagram showing ranges of susceptibilities for various sedimentary rocks. Reference sources are 1. Telford and others (1976), 2. Carmichael (1982), 3. Fedynskiy (1967), and 4. Grant and West (1965).

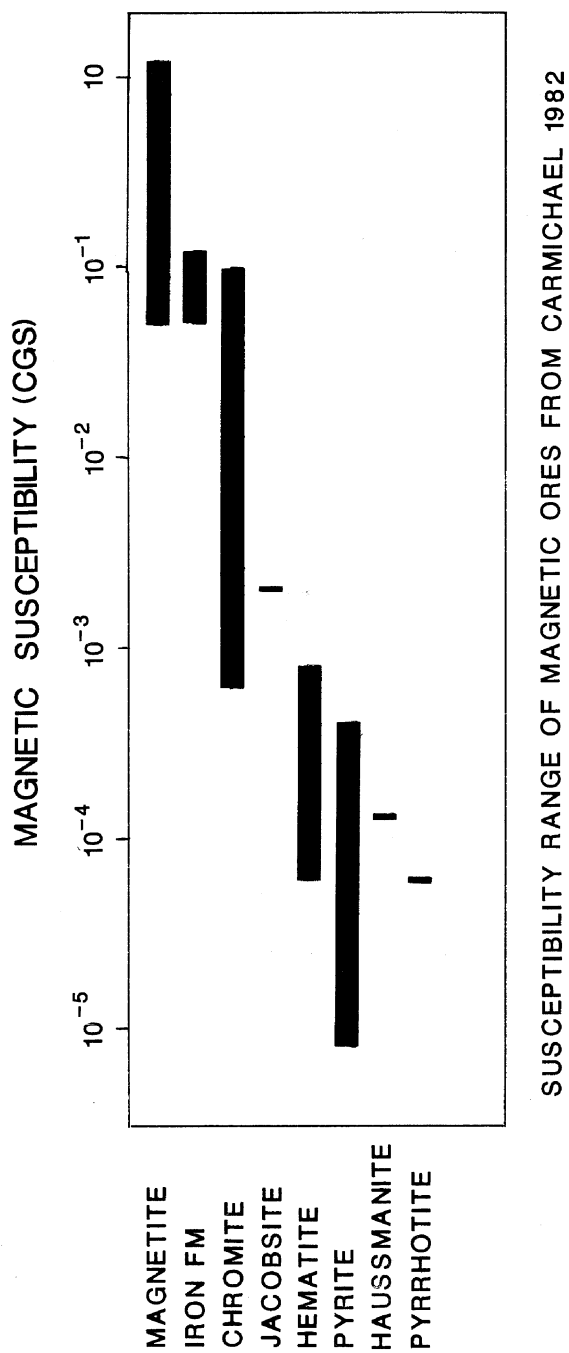


Figure 13. Diagram showing ranges of susceptibilities for eight different ores. Reference sources are Carmichael (1982) and Parasnis (1966).

Table 4. Selected values of the Koenigsberger ratio Q for various rock types from Carmichael (1982) and Kuz'micheva and Diomidova (1968).

| <i>Rock type</i> | <i>Carmichael range of Q</i> | <i>Kuz'micheva and Diomidova mean value</i> | <i>range of Q</i> |
|------------------------------|---|---|--------------------------------|
| Marine sedimentary and shale | | 5. | |
| Siltstone | 0.02 - 2.0 | | |
| Sandstone | 1.0 - 4.4 | | |
| Granite | 0.1 - 1.0 | | 0.3 - 3.4 |
| Granite (Oklahoma) | | 28. | |
| Granodiorite | 0.1 - 0.2 | | 0.3 - 3.2 |
| Dolerite | 2.0 - 3.5 | | |
| Diobase | 0.2 - 3.5 | | |
| Gabbro | 1.0 - 9.5 | | 0.3 - 3.05 |
| Volcanics (unspec.) | 30. - 50. | | |
| Basalt | 1.0 - 160. | | |
| Magnetite ore | 1.0 - 94. | | |
| Quartz-diorite | | | 0.3 |
| Granosyenite | | | 0.6 - 1.1 |

exploration. Domenico (1984) has suggested that the V_p/V_s ratio (γ) and Poisson's ratio (σ) given by

$$\sigma = \frac{0.5(V_p/V_s)^2 - 1}{(V_p/V_s)^2 - 1} \quad (8)$$

is characteristic of some sedimentary lithologies with sandstones having Poisson's ratio in the range of 0.17-0.26, dolomites 0.27-0.29 and limestones 0.29-0.33. Thus seismic methods may have potential for lithologic discrimination between sedimentary rocks. Tatham (1982), however, concludes that crack and pore geometry are the prime determinants of the V_p/V_s ratio. Fracture porosity and degree of water saturation in hard rocks also are primary causes for a lowering of both P-wave and S-wave velocity in the near surface. It is not until pressures of about 1 kilobar are obtained that fractures close sufficiently for the seismic velocities to be representative

of the true matrix velocity of the rock. Thus, in-situ velocities of ore and host rocks at most mining depths are dependent on the extent of fracturing.

Several authors have related seismic p velocity to density (ρ) (see Grant and West, 1965; Gardner and others, 1974; or Fedynskiy, 1967). The relationships between p velocity and density given in the above references are shown in figure 14. The relationship given by Gardner and others (1974) is

$$V_p = 357.4\rho^6 \quad (9)$$

while Fedynskiy (1967) gives

$$V_p = a\rho - b \quad (10)$$

where a and b are constants given as a = 6, b = -11 (Puzyrev) or a = 7.5, b = 14.8 (Urupov). Within the density range 2.4 to 3.0 gm/cm³ the velocity functions are all approximately linear.

Woeber and others (1963) provide additional information on the P and S velocity anisotropy and density of a variety of mostly igneous rocks.

A number of papers (Costagna and others, 1985; Domenico, 1984; and Pickett, 1963) empirically relate velocity (V) and porosity (ϕ) for clastic rocks. Pickett proposed the rather simple relationship

$$1/V = A + B\phi \quad (11)$$

where A+B are constants.

Faust (1953) obtained an empirical relationship dependent on the depth of burial D and the formation resistivity (ρ) in the form:

$$V = C (D\rho)^{1/6} \quad (12)$$

where C is a constant. However, there is large scatter in the data fitted.

It is doubtful that these empirically determined relationships can be extended to low-porosity igneous or metamorphic rocks.

A theoretical relationship between thermal properties and velocity has also been developed. This will be covered in the section on thermal properties.

In minerals exploration some or all of the host and cover rocks may be in the vadose zone. Under these conditions the degree of water saturation will affect the sonic velocity even of low porosity rocks. Table 5 gives the variation between saturated and air dry samples of igneous and metamorphic rocks and their porosity summarized from Christensen (1982). As can be seen, the variations can be significant for these low porosity examples.

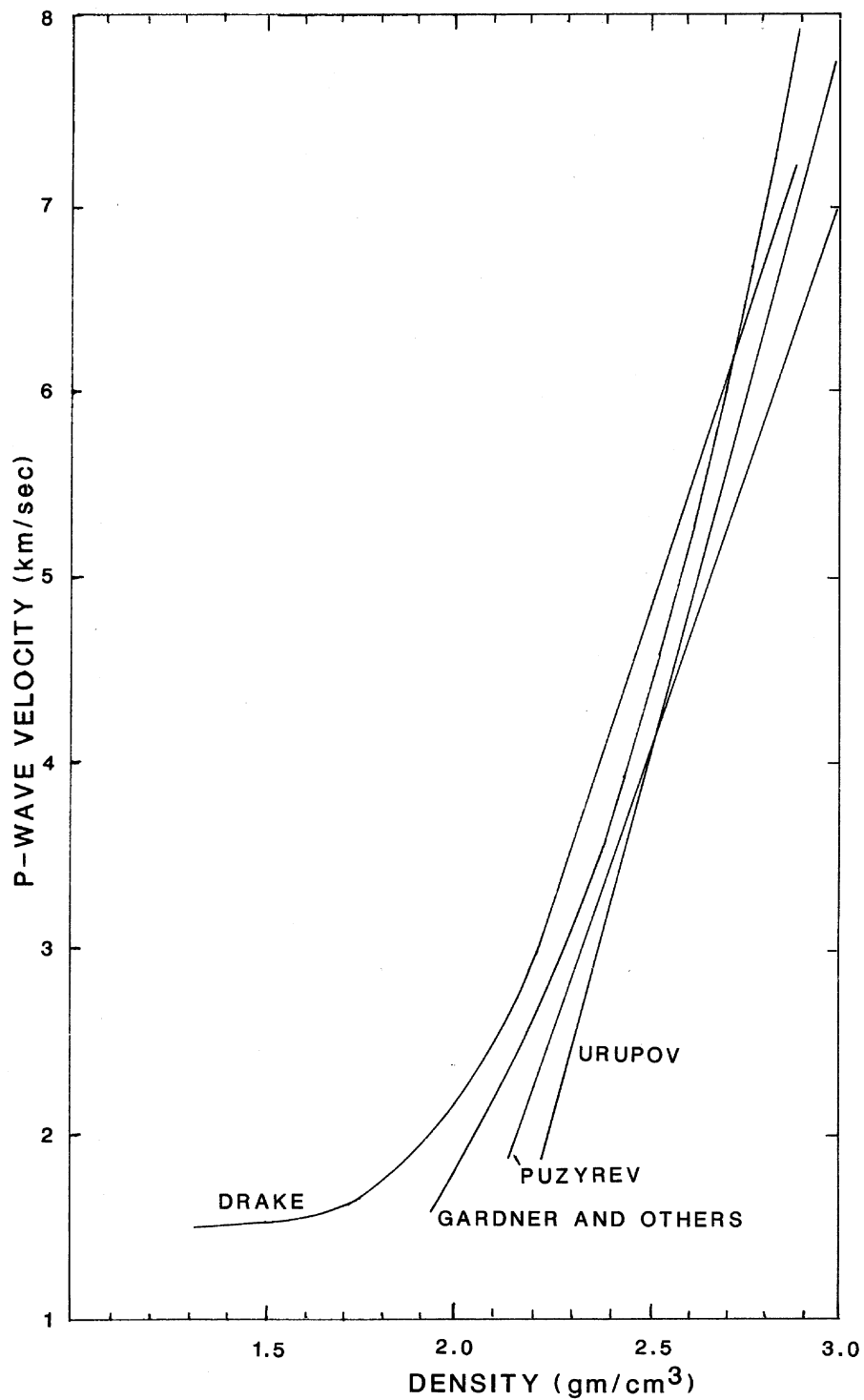


Figure 14. Graph showing empirical relationships between seismic P-wave velocity and rock bulk density from 1. Drake (Grant and West, 1965); 2. Gardner and others (1974); 3. Puzyrev (Fedynskiy, 1967); and Urupov (Fedynskiy, 1967).

Figure 15 illustrates the ranges of P-wave velocities for igneous and metamorphic rocks from several sources. Values shown from Press (1966) are at a confining pressure of 10 bars. Christensen (1982) gives values for a wide variety of igneous and metamorphic rocks, but has not provided a summary. However, in figure 15 the range for peridotite has been extracted, with the low values representative of serpentinized peridotite and higher values correspond to less altered rock. Alteration of other igneous or metamorphic rocks would be expected to similarly lower the seismic velocity.

Figure 16 presents the velocity range associated with sediments and sedimentary rocks. Unconsolidated sediments are seen to have very low velocities.

Table 6 lists seismic velocity ranges of ores or ore minerals from several sources. Note the low velocity of graphite. Jadeite has the highest velocity in the listing, not surprisingly being similar to that of pyroxenite in table 5.

Table 5. Seismic P-wave velocities for selected igneous rocks as a function of water saturation (adopted from Christensen, 1982).

| <i>Rock type</i> | <i>porosity range %</i> | <i>Vp range air dry km/sec</i> | <i>Vp range saturated km/sec</i> |
|------------------|-----------------------------|------------------------------------|--------------------------------------|
| Granite | 0.6 - 1.8 | 3.25 - 5.40 | 5.1 - 6.3 |
| Syenite | 0.7 - 1.1 | 4.10 - 5.45 | 5.40 - 6.45 |
| Gneiss | 0.3 - 0.9 | 3.65 - 5.20 | 5.45 - 6.00 |
| Granulite | 0.2 - 0.4 | 4.90 - 5.60 | 5.45 - 6.00 |
| Diorite | 0.6 - 1.2 | 4.65 - 5.90 | 5.95 - 6.6 |
| Norite | 0.2 | 6.15 - 7.00 | 6.70 - 7.10 |
| Pyroxenite | 0.2 | 7.10 - 8.15 | 7.70 - 8.40 |

ELECTRICAL PROPERTIES

Resistivity

In the upper kilometer or so where most minerals exploration takes place the resistivity of earth materials is even less representative of lithologic type than are seismic velocities. In this environment, earth resistivity is generally determined entirely by rock porosity and the conductivity of the fluid filling the pore space. Because porosity and fluid conductivity are not parameters closely related to lithologic type, especially for igneous and metamorphic rocks, empirical relationships between resistivity and lithologic type show a wide variation. Further complicating the picture are problems related to how representative are laboratory measurements on hand or core samples of in-situ resistivities. Samples may dry prior to measurement requiring dehydration with water that may not be representative of the original connate water, and small samples may not be representative of the effect of fractures in the bulk. In part because of these problems, several authorities do not even provide summary results of resistivities of-wet igneous or metamorphic rocks as a function of specific lithologic types (Keller, 1966, Keller and Frischknecht, 1966, Keller, 1982, and Olhoeft,

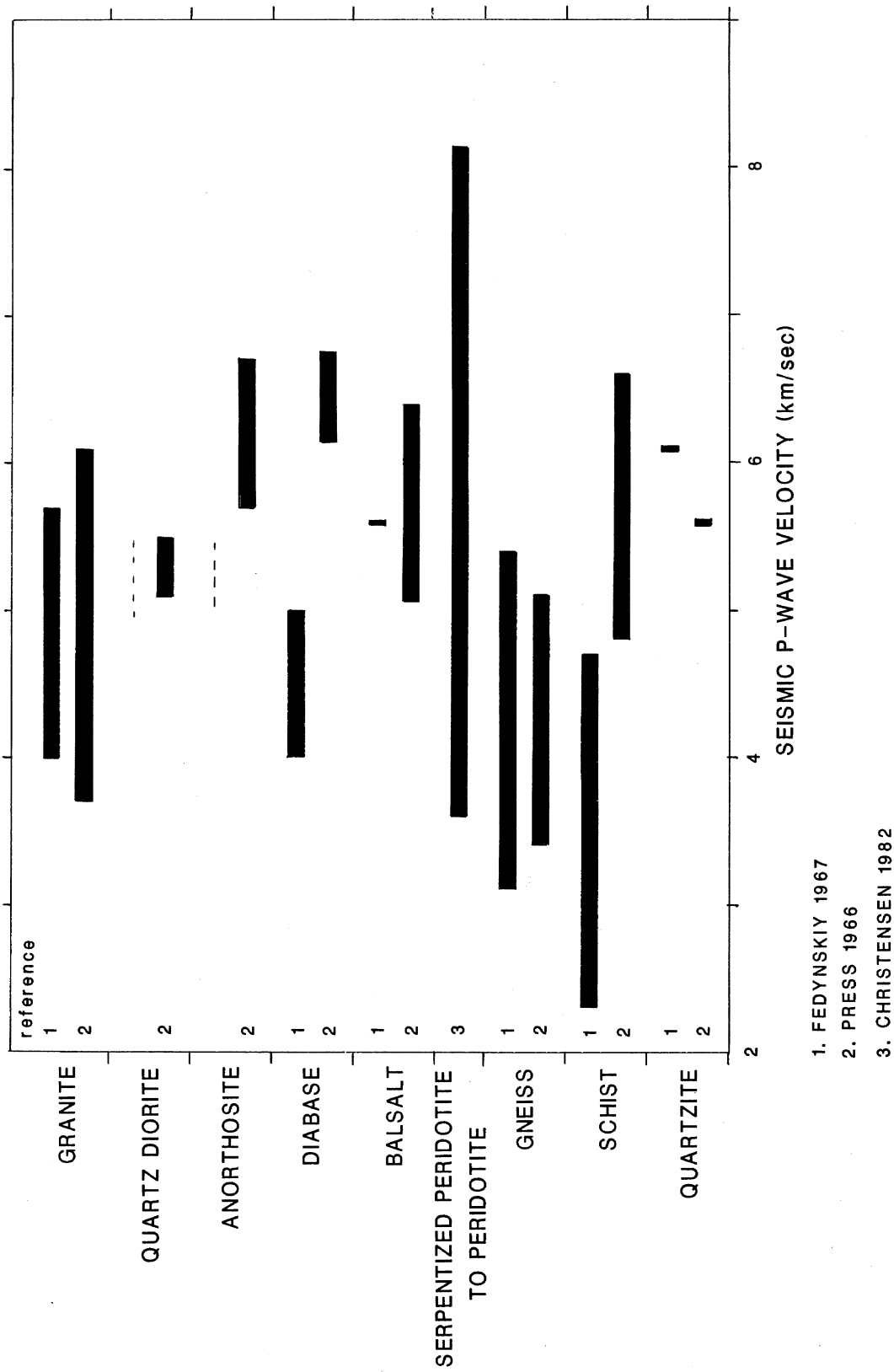


Figure 15. Range of P-wave velocities for selected igneous and metamorphic rocks. Reference sources are 1. Press (1966), 2. Fedynskiy (1967); and 3. Christensen (1982).

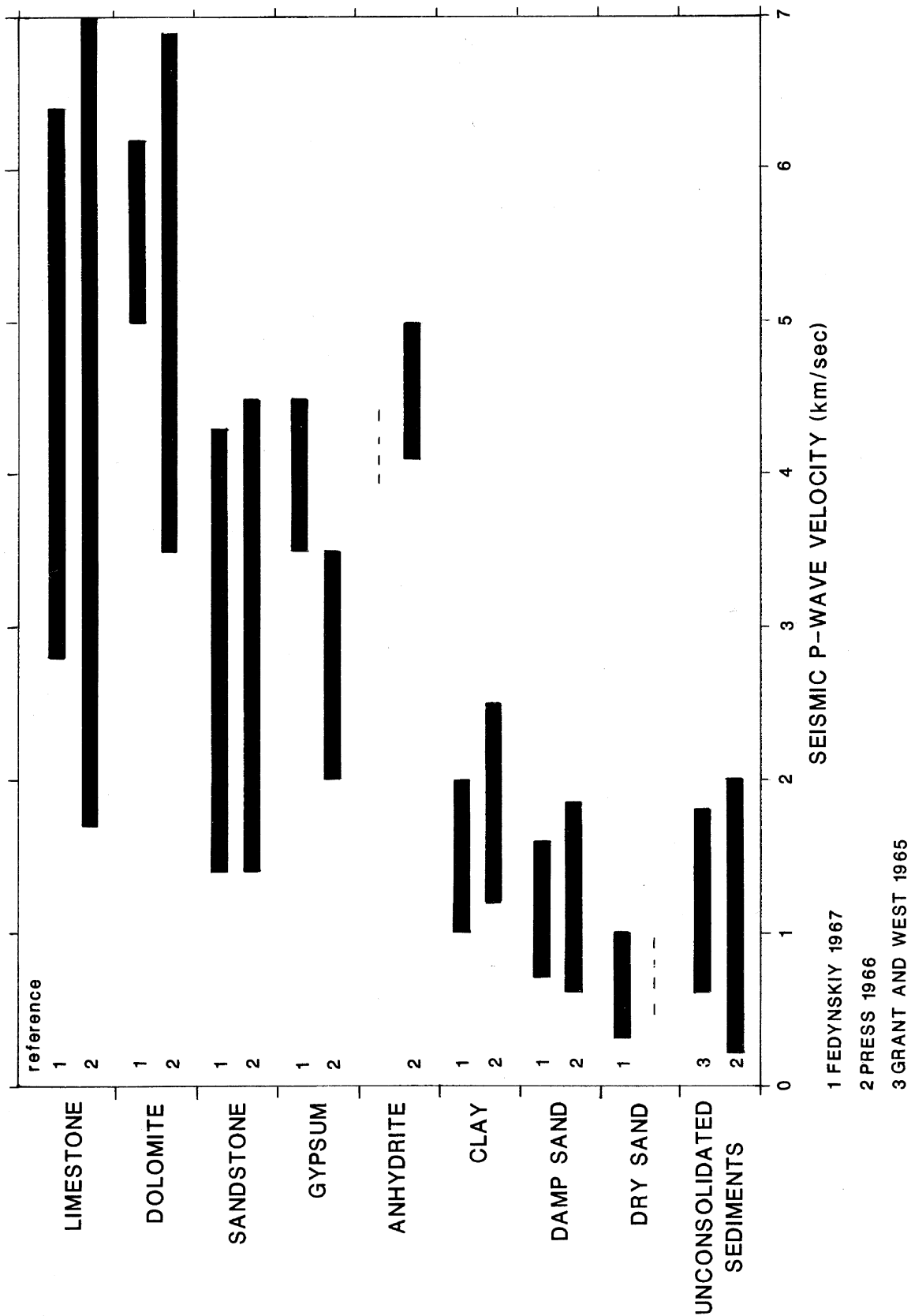


Figure 16. Range of P-wave velocities for selected sedimentary rocks. Reference sources are 1. Press (1966); 2. Fedynskiy (1967); and 3. Grant and West (1965).

1981) . Sedimentary rocks because of their importance to the petroleum industry are more extensively studied.

Table 6. *Seismic P- and S-wave velocities for selected ores and ore minerals from Woeber and others (1963) and Christensen (1982) .*

| <i>Ore mineral</i> | <i>density gm/cm³</i> | <i>Average P velocity km/sec</i> | <i>Average S velocity km/sec</i> |
|--------------------|--------------------------------------|--------------------------------------|--------------------------------------|
| graphite | 2.16 | 3.06 | 1.86 |
| limonite | 3.55 | 5.36 | 2.97 |
| pyrrhotite | 4.55 | 4.69 | 2.76 |
| pyrite | 4.81 | 7.69 | 4.78 |
| magnetite | 4.81 - 6.77 | 4.18 - 4.87 | 1.97 |
| hematite | 4.93 - 5.00 | 6.82 - 7.72 | 3.84 |
| siderite | 3.57 | 7.01 | |
| magnesite | 2.80 - 2.97 | 7.11 - 8.12 | |
| jadeite | 3.18 - 3.33 | 8.21 - 8.67 | |
| anhydrite | 2.63 - 2.93 | 4.90 - 2.29 | |
| gypsum | 2.29 | 4.95 | |
| halite | 2.16 | 4.13 | |

Illustrative of the broad range of resistivities measured on rocks and the lumping of major lithologic types are figures given by Grant and West (1965) and similar illustrations from Sumner (1976), that were reproduced by Hallof (1980). These are shown in figure 17 redrafted to the same scale for comparison. The figure shows resistivities covering a span of ten orders of magnitude, 10^{-1} to 10^5 ohm-meters. The upper four decades represent measurements on laboratory samples of low porosity or very dry samples, and are not representative of in-situ bulk resistivities which will rarely exceed 10^5 ohm-m. Again, this emphasizes the need for caution when applying laboratory results to the field situation.

Below the water table, and in rocks where electronic conduction processes due to metallic luster sulfide minerals can be neglected, conduction is controlled by migration of charged ions in the pore water. The charged ions can be considered to be particles moving under the force of an applied electric field in a viscous fluid. The current then is proportional to the number of ions, ion size, water viscosity, and electric field. Rock resistivity then is related principally to the concentration of dissolved and ionized species in solution, and water viscosity since ion size does not vary appreciably. Viscosity is a factor when the pore water temperature changes. Increasing temperature decreases the viscosity and, decreases the resistivity.

Although the connate water resistivity and porosity control the resistivity of most rocks, connate water resistivity typically does not vary

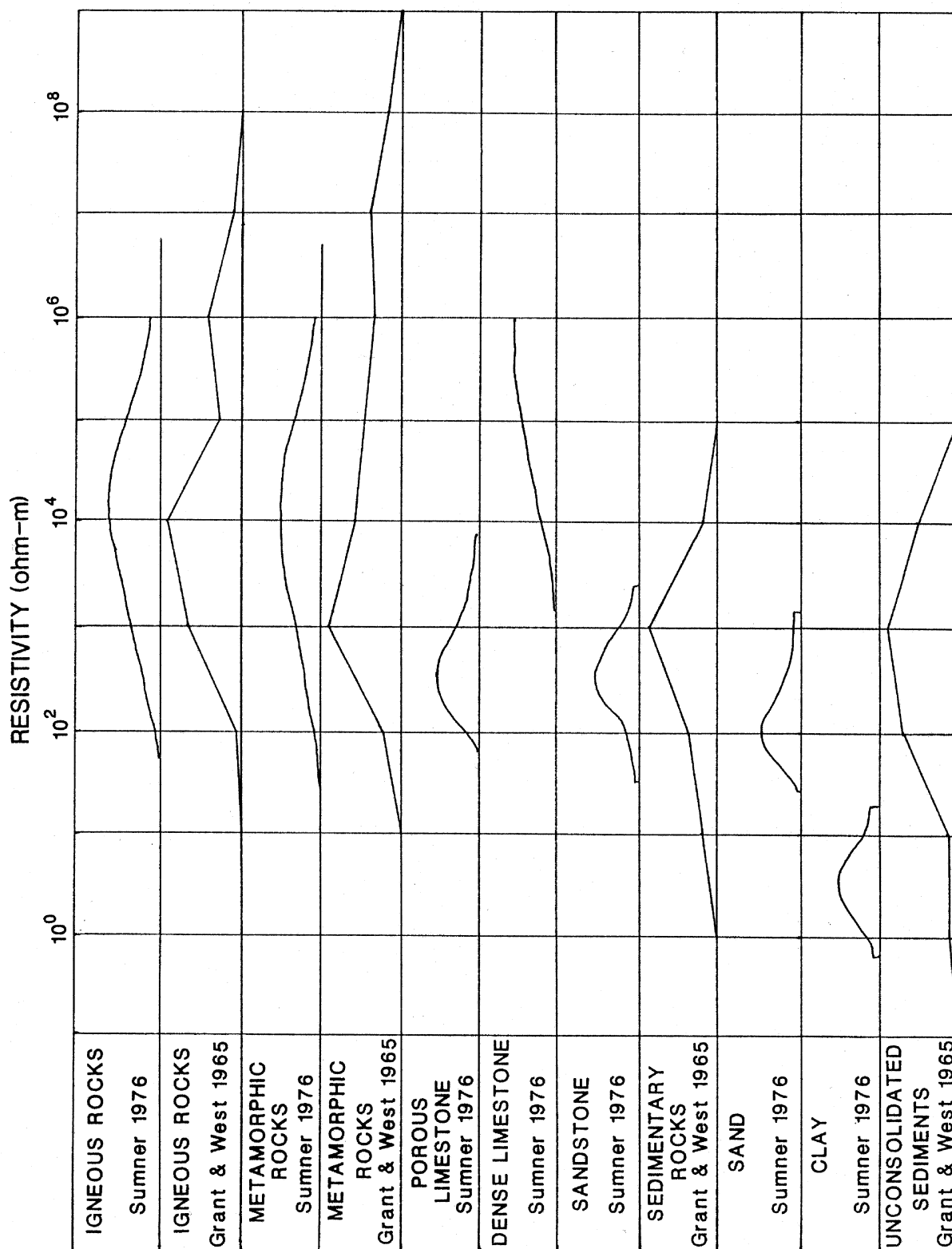


Figure 17. Distribution diagrams of resistivity values for several types of rocks from Sumner (1976) and Grant and West (1965)

over a wide range. Table 7 from Keller and Frischknecht (1966), gives average values of connate water resistivity observed for a number of regions and lithologies. In general, shallow waters and waters from crystalline rocks have the highest resistivities, while deeper waters, or those from sediments, show the lowest resistivities. The average resistivity range is seen to be a little over two orders of magnitude.

The more extensive studies of sedimentary rocks have resulted in an empirical relationship for sediments called Archie's law that relates resistivity (ρ) to porosity (ϕ) and the connate water resistivity (ρ_w). The relationship is:

$$\rho = A\rho_w\phi^{-m} \quad (13)$$

where A and m are empirical constants dependant on the type of sediment. Keller and Frischknecht (1966) give values of A and m for a number of sediments in which A varies from 0.47 to 2.3 and m from 1.64 to 0.23. Where reasonable estimates of sedimentary host or cover rock porosity and connate water resistivity can be made, Archie's law can be used to provide resistivities for modeling.

In the vadose zone, rock resistivities do not increase in direct proportion to the degree of undersaturation. Nonlinear effects result from the presence of a continuous film of water between grains at larger values of partial saturation, and from surface conduction effects. Keller and Frischknecht (1966) provide further details on these effects.

Desert soils and alluvium exhibit an interesting paradox in the vadose zone where the dryness might be expected to result in increased resistivity over that of soils in moister regions. However, most desert soils show resistivities in the range of 10-100 ohm-mg about an order of magnitude lower than soils from regions with average rainfall. This is due to the compensating effect of increased salt content in the residual moisture held by the soils.

| Table 7. Connate water average resistivities from various regions and lithologies (from Keller and Frischknecht, 1966). | |
|---|--|
| <i>Water Source</i> | <i>Average Resistivity (ohm-m)</i> |
| European igneous rocks | 7.6 |
| South African igneous rocks | 11.0 |
| South African metamorphic rocks | 7.6 |
| Precambrian Australian metamorphic rocks | 3.6 |
| Pleistocene to recent European sedimentary rocks | 3.9 |
| Pleistocene to recent Australian sedimentary rocks | 3.2 |
| Tertiary European sedimentary rocks | 1.4 |
| Tertiary Australian sedimentary rocks | 3.2 |
| Mesozoic European sedimentary rocks | 2.5 |

| Table 7. Connate water average resistivities from various regions and lithologies (from Keller and Frischknecht, 1966). | |
|---|-------|
| Paleozoic European sedimentary rocks | 0.93 |
| Oil field chloride waters | 0.16 |
| Oil field sulfate waters | 1.2 |
| Oil field bicarbonate waters | 0.98 |
| Jurassic Colorado-Utah USA waters | 1.8 |
| Cisco series Texas, USA. waters | 00061 |
| Pennsylvanian, Oklahoma, USA waters | 0.062 |

For a given rock type the resistivity normally increases with age and depth of burial, principally because of the associated decrease in porosity. Keller and Frischknecht (1966) give a table of resistivity ranges for rocks of various ages and types that shows a typical change of 1.5 to 2 orders of magnitude increase from Quaternary to Precambrian for the same type of rock. These results are shown graphically in figure 18. The five orders of magnitude range shown here is more representative of in-situ values measured in the earth.

The presence of electronic conductors, such as graphite and the metallic luster sulfides, do not normally contribute to the lowering of host and cover rock resistivities. Their presence as ore or in association with ore is often what makes an ore body detectable by electrical methods. These situations are covered in the individual models under properties of the deposit itself. Occasionally, a situation is encountered where sulfides or graphite may be deposited as thin continuous films in a network along the fracture planes of rocks. Stockwork deposits are a good example of where this may occur. When this happens, a few percent of these electronic conductors can cause a dramatic lowering of resistivity suggestive of much greater mineral content. Graphite is particularly notable for its ability to reduce rock resistivity when present in small quantity along fracture planes (Grant and West, 1965).

Although the above discussion identifies many problems that appear to make the application of resistivity methods difficult in minerals exploration, in practice there are many straightforward uses. Mafic rocks are found to weather to lower resistivity colluvial material than felsic rocks. This permits mapping of such units by airborne electromagnetic methods as demonstrated by Palacky (1986) in Brazil, and to exploration for kimberlites pipes (Gerryts, 1967). Hydrothermal ore deposits often have large alteration haloes extending for some distance into various host rocks. Alteration processes often increase the porosity of the host rock during argillization and propylitization, or if silicification occurs, the porosity may decrease. Either of these processes can provide a clear resistivity contrast that can be detected. Deposits often are controlled by fracturing and faulting which causes increased host rock porosity that can be detected by the associated lower resistivity. However, in practice, within given districts, lithologic units often will have uniform electrical properties that do permit them to be mapped.

Sedimentary and metasedimentary rocks, especially those that show a strong orientation of pore spaces, or that contain interbedded units, may exhibit significant resistivity anisotropy. Keller and Frischknecht (1966)

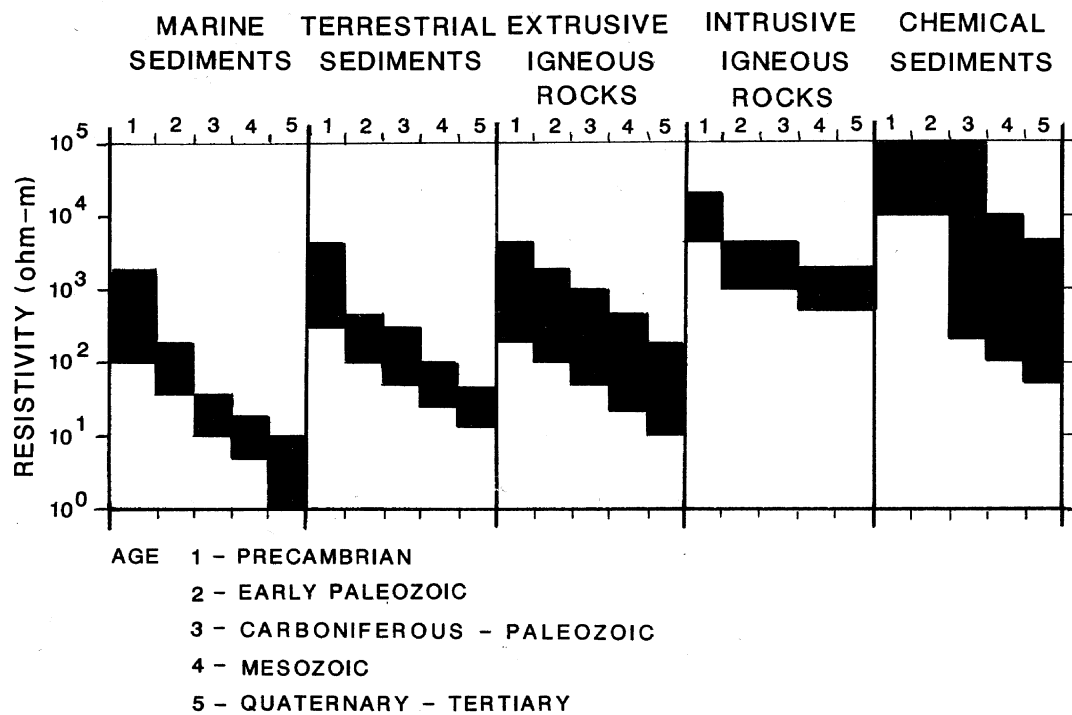


Figure 18. Diagram showing variation of resistivity as a function of age for marine and terrestrial sediments, extrusive and intrusive igneous rocks and for chemical sediments (from Keller and Frischknecht, 1966).

give values of coefficients of anisotropy for various rock types, some of which show resistivity differences exceeding a factor of 10. Crystalline rocks in which a preferred fracture orientation exists also may show significant anisotropy.

Figure 19 shows ranges of resistivities for nine types of sedimentary rocks given by Fedynskiy (1967) and Telford and others (1976). This figure can be compared with results for similar sedimentary rocks given in figure 17.

Figure 20 shows resistivity ranges for twelve types of igneous and metamorphic rocks also taken from Fedynskiy (1967) and Telford and others (1976). On average, these rock types show a wide range of overlap in values.

Telford and others (1976) also give a listing of resistivities observed on a variety of metallic ores, to which the interested reader is referred.

IP Effect

The presence of polarizable minerals, principally metallic-luster sulfides, graphite, clays, and zeolites, within a rock mass are measured by the induced polarization method. Several electrochemical processes, occurring within the earth during passage of electrical current, have been proposed as the cause for the IP phenomenon (Keller and Frischknecht, 1966; Sumner, 1976; Telford and others, 1976; Hallof, 1980). However, some aspects of the conventional theory have recently been questioned (Fink, 1979). But, no matter what the details of the source mechanisms, the IP method is a very effective tool in exploration where sulfides are present. This is especially the case for disseminated sulfides because the polarization phenomenon is proportional to the active surface area of the sulfides within the pore spaces of the host rock.

The polarization response for sulfides is a complex function of a number of parameters including, host rock porosity, grain size, degree of water saturation, current density, and sulfide content. Sumner (1976) states that sulfide content as low as 0.5% may be detected. Membrane polarization response, related to clay or zeolite content, is dependent on the type of clay or zeolite, the cation exchange capacity, and how the minerals are arranged within the pore spaces of the host rock. Keller and Frischknecht (1966) give a qualitative explanation for this type of polarization phenomenon, and show that maximum polarization occurs for clay content generally less than 10% of pore volume. In the case of montmorillonite, the maximum occurs near 0.5% of pore volume. Readers should consult the references cited for details of these complex phenomena.

Equipment used to measure the polarization response of rocks operates in either the time or frequency domain, with several variations of these two methods in use. No measurement standards are in place, and due to instrumentation differences a variety of parameters have been used to measure and report the magnitude of polarization response. These various parameters are not dimensionally equivalent. The more common parameters are frequency effect (FE) or percent frequency effect (PFE), chargeability (M), phase angle (ϕ), and metal factor (MF). To add to the confusion, the above terms are not defined consistently. For example, the frequency effect (FE) which is the normalized difference in resistivity measured at two separate frequencies has been defined as:

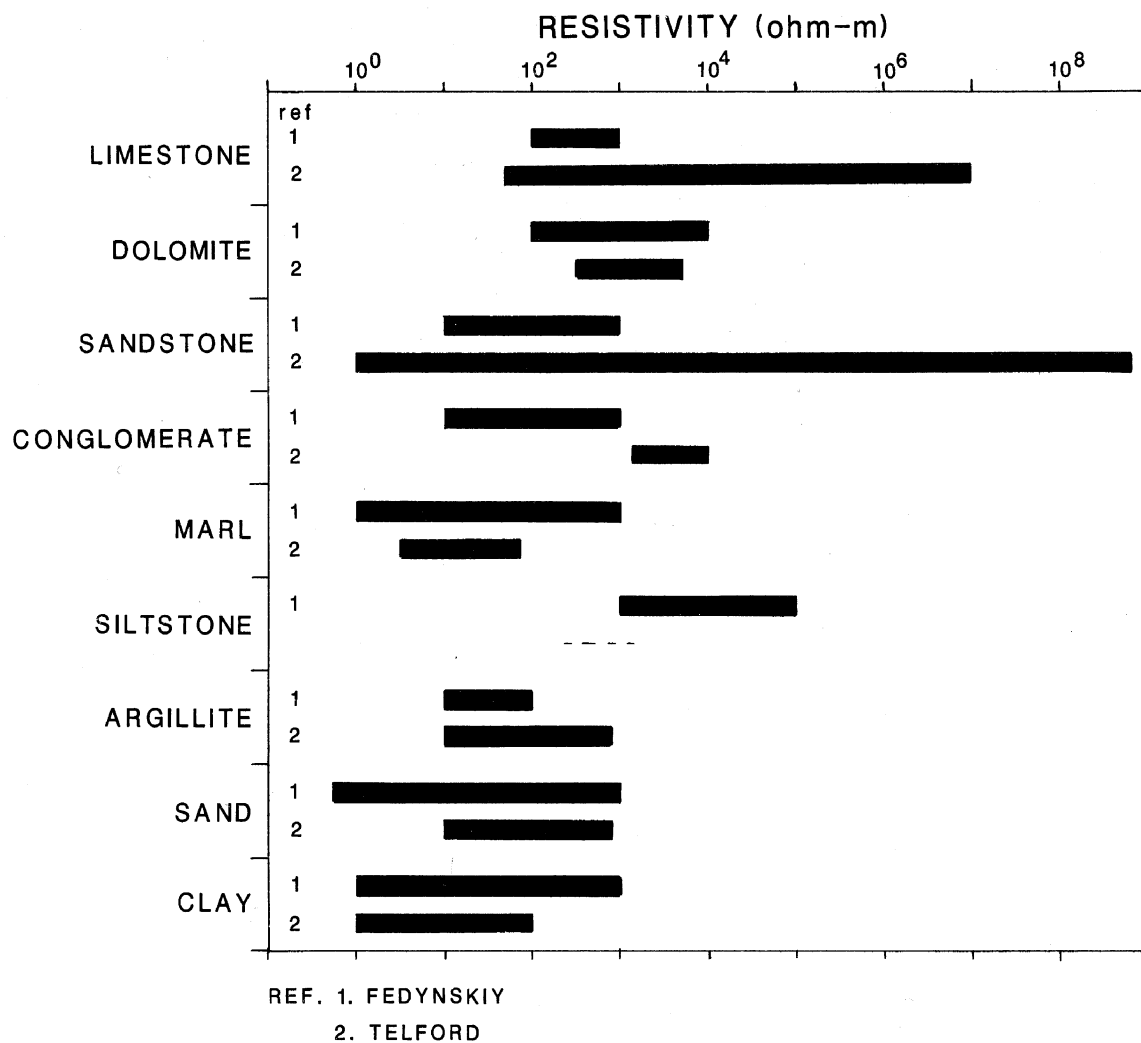


Figure 19. Ranges of resistivities for selected sedimentary rocks from Fedynskiy (1967) and Telford and others (1976).

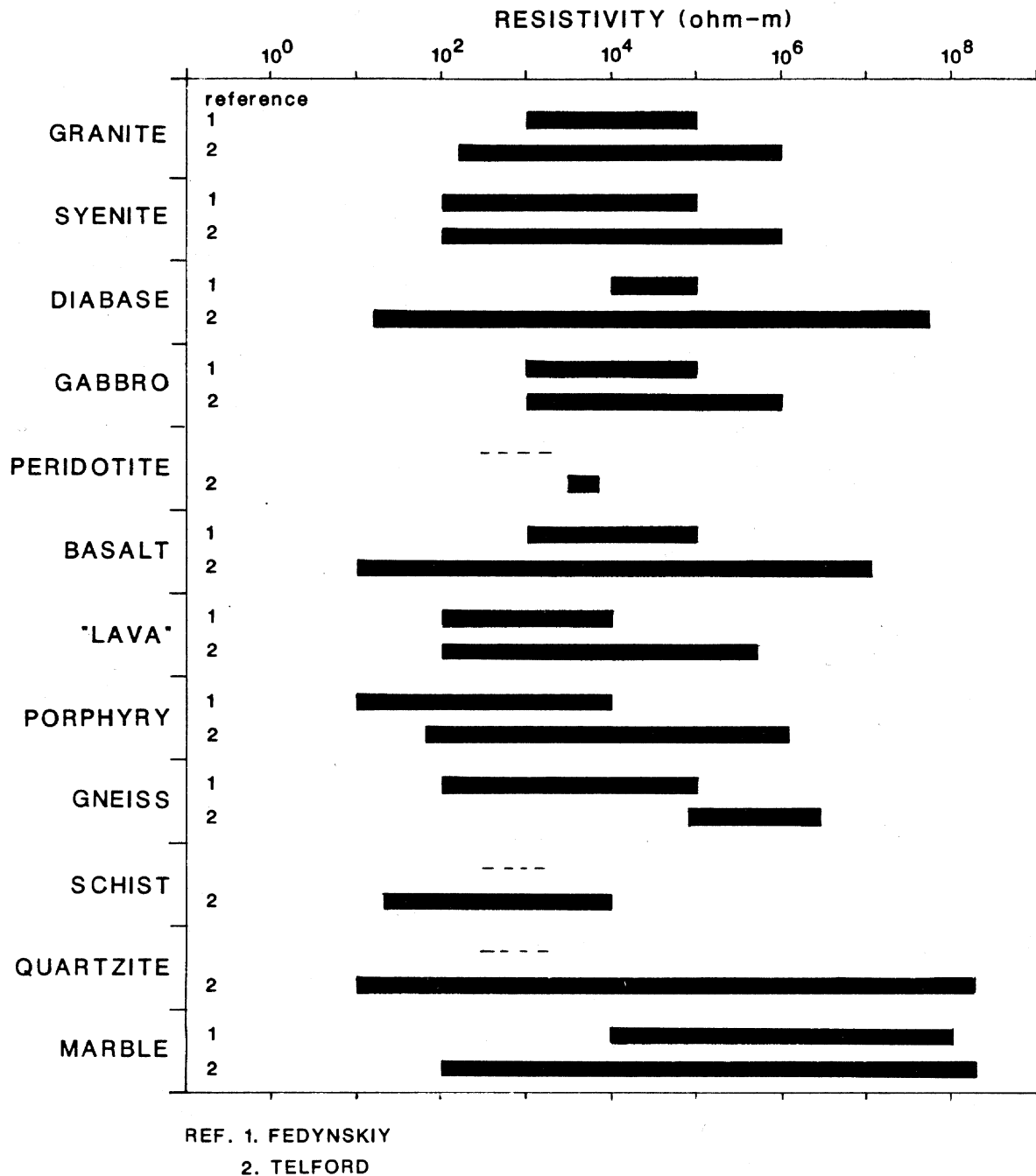


Figure 20. Ranges of resistivities for selected crystalline rocks from Fedynskiy (1967) and Telford and others (1976).

$$1. \text{ FE} = \frac{Q_{f1} - Q_{f2}}{Q_{f2}} \quad \text{Telford and others, 1976; Sheriff, 1991; Rogers, 1966; Madden and Marshall, 1959} \quad (14)$$

$$2. \text{ FE} = \frac{Q_{f1} - Q_{f2}}{Q_{f1}} \quad \text{Brant and others, 1966} \quad (15)$$

$$3. \text{ FE} = \frac{Q_{f1} - Q_{f2}}{\sqrt{Q_{f1} Q_{f2}}} \quad \text{Keller and Frischknecht, 1966.} \quad (16)$$

where frequency $F_2 > F_1$

Two or more frequencies may be measured, but should always be specified.

The metal factor (MF) or metal conduction factor has been defined as

$$4. \text{ MF} = (2\pi \cdot 10^5) \frac{Q_{f1} - Q_{f2}}{Q_{f1} Q_{f2}} \quad \begin{array}{l} \text{Telford and others, 1976; Brant} \\ \text{and others, 1966;} \\ \text{Madden and Marshall, 1959} \end{array} \quad (17)$$

$$5. \text{ MF} = \frac{10^5 \text{ FE}}{Q_{f1}} \quad \text{Rogers, 1966.} \quad (18)$$

$$6. \text{ MF} = \frac{2000 \text{ PFE}}{Q_a} \quad (Q_a \text{ not specified at } f_1 \text{ or } f_2) \text{ Sumner, 1979} \quad (19)$$

$$7. \text{ MF} = 10^3 \times \frac{M(\text{chargeability})}{Q_a} \quad \text{Witherly and Vyselaar, 1990} \quad (20)$$

In the time domain, chargeability is commonly measured, and this has been defined as simply the ratio of the primary voltage (V_p) to the secondary (decay) voltage (V_s) measured at a specific time after current shut-off and is dimensionless.

$$8. M = \frac{V_s}{V_p} \quad \begin{array}{l} \text{Brant, 1966; Telford and others, 1976;} \\ \text{Sheriff, 1991; Sumner, 1976} \end{array} \quad (21)$$

The secondary voltage may be measured at one or several times during the decay.

Chargeability has also been defined as the time integral of the voltage decay curve between two specified times normalized to the primary voltage

$$9. M = \frac{1}{V_p} \int_{\tau_1}^{\tau_2} V_{\tau} dt \quad \text{Sheriff, 1991; Sumner, 1976.} \quad (22)$$

Historically, frequency domain measurements were reported in FE or PFE, but more recently instrumentation has been developed to give the amplitude and phase angle ϕ between transmitted and received voltage as a measure of polarization response. Typically phase response is reported in multiradians at a particular frequency. When such instrumentation is used over a wide frequency range the method is referred to as spectral IP or the complex-resistivity method. This method provides far more information on electrical relaxation phenomena (information on electrical relaxation phenomena may be found in Pelton and others (1978) and Olhoeft (1981)) in the earth than more convention IP techniques in which measurements are taken at only two or three frequencies or time delays. The increased information permits, in some cases, discrimination between polarizable minerals present in the rock. Sumner (1976), Pelton and others (1978) and a series of papers in Fink and others (1990) provide details on the complex resistivity method.

Although theoretical and empirical relationships between the various polarization measures have been given, comparisons between published results are often impossible because the frequencies or times over which measurements were made are often not stated. Thus, compilations of physical properties such as presented here become impractical because reported IP parameters cannot be made commensurate. Part of the problem is that IP investigators have been reporting the response measured by individual instruments, rather than a direct measure of a rock property. A step towards reporting of a direct rock property is calculating of the Cole-Cole parameters derived from IP measurements (Pelton and others, 1978). However, a field test comparing Cole-Cole parameters derived from frequency domain and time domain systems (Johnson, 1990) showed correspondence between areas of high and low-values, but relatively poor quantitative agreement.

Thus, for cover rock polarizability properties we do not present comparisons between tables derived from various sources. In figure 21 we give ranges of IP response presented as metal factor for several types of lithologies and ores from Madden and Marshall (1959) which were derived from both laboratory and field work. The definition of MF is as given in equation 17 at frequencies of 10 Hz and "dc". This figure serves to show the difference between lithologies and relative range of values of this particular

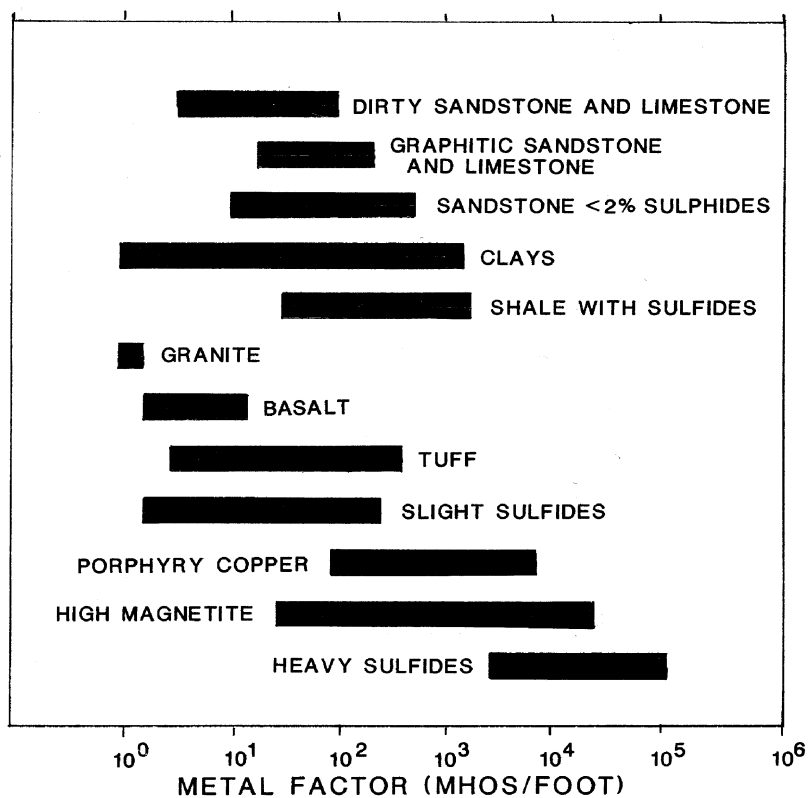


Figure 21. Range of IP response shown as metal factor for several types of rocks and ores from laboratory and field measurements from Madden and Marshall (1959).

polarization measure. Telford and others (1976) also give several tables showing typical polarization responses of geologic materials. When using the accompanying models descriptions, the problems of measures of polarization response need to be kept in mind. Where quantitative values are quoted, the reader wishing to use these should go to the original reference for clarification.

Electrokinetic coupling coefficient

Self potentials arising from current flow through a natural electrochemical cell in the earth, related to oxidation-reduction reactions are not directly related to a specific rock property. Thus, no listing of properties can be given relevant to this type of self potential source. Also, water moving through the earth will generate a self potential signal by an electrokinetic mechanism. Normally, voltages generated by the process would be noise that perturbs the signal from a sulfide deposit. However, water flow controlled by fractures and faults may generate a self potential field than will define the structure. The self potential generated is determined by the electrokinetic coupling coefficient of the rocks. Johnson (1983) gives laboratory values for a variety of rock types, and summarizes results for various groups of rocks. His summary results are shown in figure 22 for the mean of each group and one standard deviation to each side.

Johnson (1983) in his summary did not include rocks of high permeabilities which also showed very high coupling coefficients. Values of 34 to 226 mv/atm. were observed on samples of pegmatite, pumice, tuff, scoria, tuffa, and coquina. Only two marble samples, and one limestone sample gave measurable negative values.

OPTICAL PROPERTIES

Spectral Reflectance

Spectral reflectance is a physical property of materials that describes how light in a continuous electromagnetic spectrum interacts with the material. In simple terms, incident light in the optical portion of the spectrum (visible and near-infrared) can be: (1) transmitted, (2) absorbed, and (3) reflected such that,

$$\tau + \alpha + \rho = 1, \quad (23)$$

where

τ = transmissivity,
 α = absorptivity, and
 ρ = reflectivity.

For nearly all naturally occurring materials at the earth's surface, the transmissivity is 0, so that,

$$\alpha + \rho = 1. \quad (24)$$

Absorption and reflection are sensitive to the wavelength of the incident light, that is, reflectivity and absorptivity are spectral properties. The amount of light reflected at any given wavelength ($\rho\lambda$) is a function of the elemental content and molecular structure of the materials, and measurement of reflected light across the visible and near-infrared

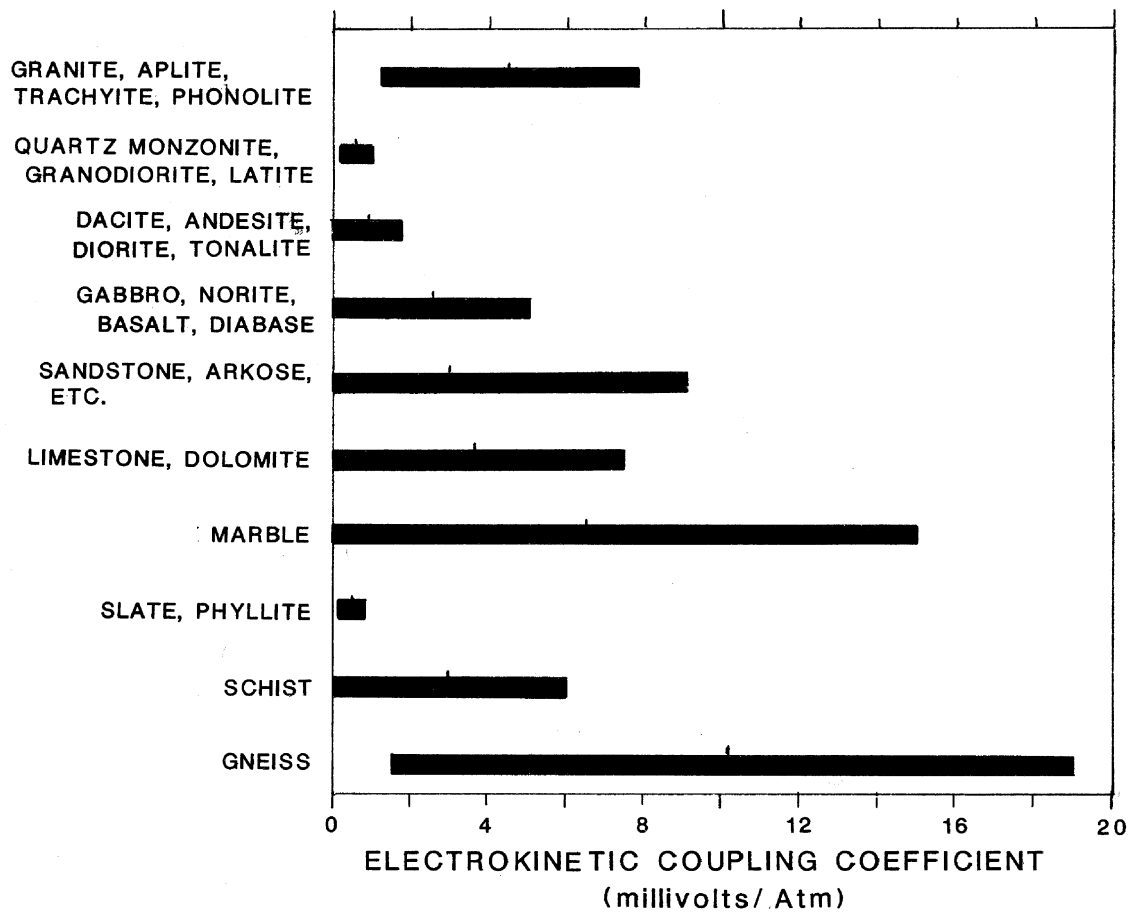


Figure 22. Electrokinetic coupling coefficients for selected lithologies from Johnson (1983). The tick indicates the mean for each lithology, and the bar extends one standard deviation on each side.

spectrum can provide information diagnostic of minerals species. Spectral curves (figure 23) show the proportion of reflected to incident light as a function of wavelength. The wavelength of reflection minima and their size and shape provides the key information for identifying minerals from their spectral reflectance characteristics. An excellent detailed discussion of electromagnetic radiation and its interaction with matter is contained in Hunt (1980).

Broad band remote sensing systems, such as the experimental airborne Thematic Mapper Simulator (TMS) and the commercially available Landsat Thematic Mapper (TM) satellite system, provide local to regional geographic coverage. But the relatively broad bands of these systems only permit the detection of broad mineral groups with similar basic spectral reflectance characteristics, such as iron oxides, hydroxyl-bearing silicates (clays and micas), and carbonates (Hunt and Salisbury, 1970, 1971; Hunt, Salisbury, and Lenhoff, 1971a, 1971b, 1972, 1973). Many of the minerals in these mineral groups are found in hydrothermally altered rocks or are weathering products derived from altered rocks (Hunt and Ashley, 1979; also see Selected Bibliography in Knepper, 1989). Broad band, multispectral imaging systems provide digital data suitable for detecting and mapping the presence and gross geometry of hydrothermal systems, as well as the distribution of selected industrial minerals, such as carbonates, clays, and zeolites.

High resolution, narrow band multispectral imaging systems, such as NASA's experimental AVIRIS (Airborne Visible and Infrared Imaging System) and the commercially available GER (Geophysical Environmental Research) instrument, obtain digital image data for each ground picture element (pixel) with a spectral resolution nearly matching the resolution of laboratory instruments. These data permit the identification of the presence of many mineral species, providing detailed information on possible mineral zoning and the occurrence of specific industrial mineral products.

THERMAL PROPERTIES

Thermal conductivity and inertia

The thermal properties of primary concern to the geophysicist are thermal conductivity, K , and thermal inertia, I . These are not independent properties, but are related by the following equation:

$$K = I^2/\rho c \quad (25)$$

where c is the specific heat and ρ is the density. Thus, rocks with high thermal conductivity will also have high thermal inertia. In this section, thermal conductivities only will be given. Those wishing values of thermal inertia may calculate them from the equation given above using densities given previously. The specific heat for rock forming minerals is nearly constant at 0.2 cal/gm °C. The principal factor complicating the relation given in equation 25 is the presence of pore water in the more porous sediments where specific heat may approach that of water at 1.0 and conductivity may increase as much as eight times. The effect of pore water on these thermal properties is evident in table 8 which gives the ranges of thermal conductivities observed for various sedimentary rocks and unconsolidated materials. Figure 24 gives thermal conductivities measured for various types of rocks. This figure shows the distinct difference between dolomite and limestone, and the high values and large range characteristic of quartzite.

A. REFLECTANCE CURVES

B. BASIC CURVE SHAPES

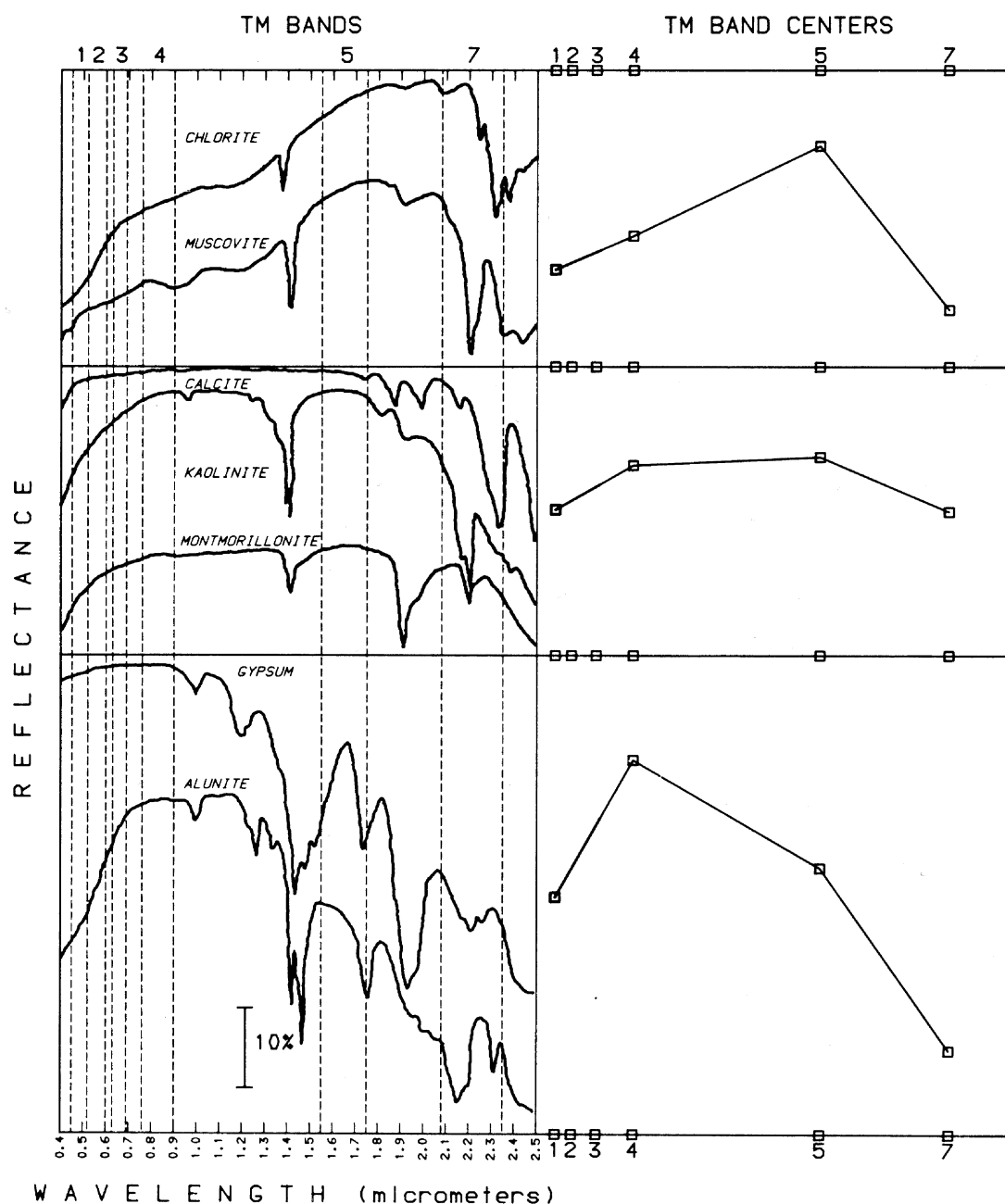


Figure 23. A. Spectral curves of common minerals often associated with hydrothermally altered rocks, showing the locations of the Landsat Thematic Mapper spectral bands. The curves are offset vertically to allow curve stacking. From R. Clark (U.S. Geological Survey, unpublished data).

B. A method of grouping the minerals based on the basic shape of their reflectance curves.

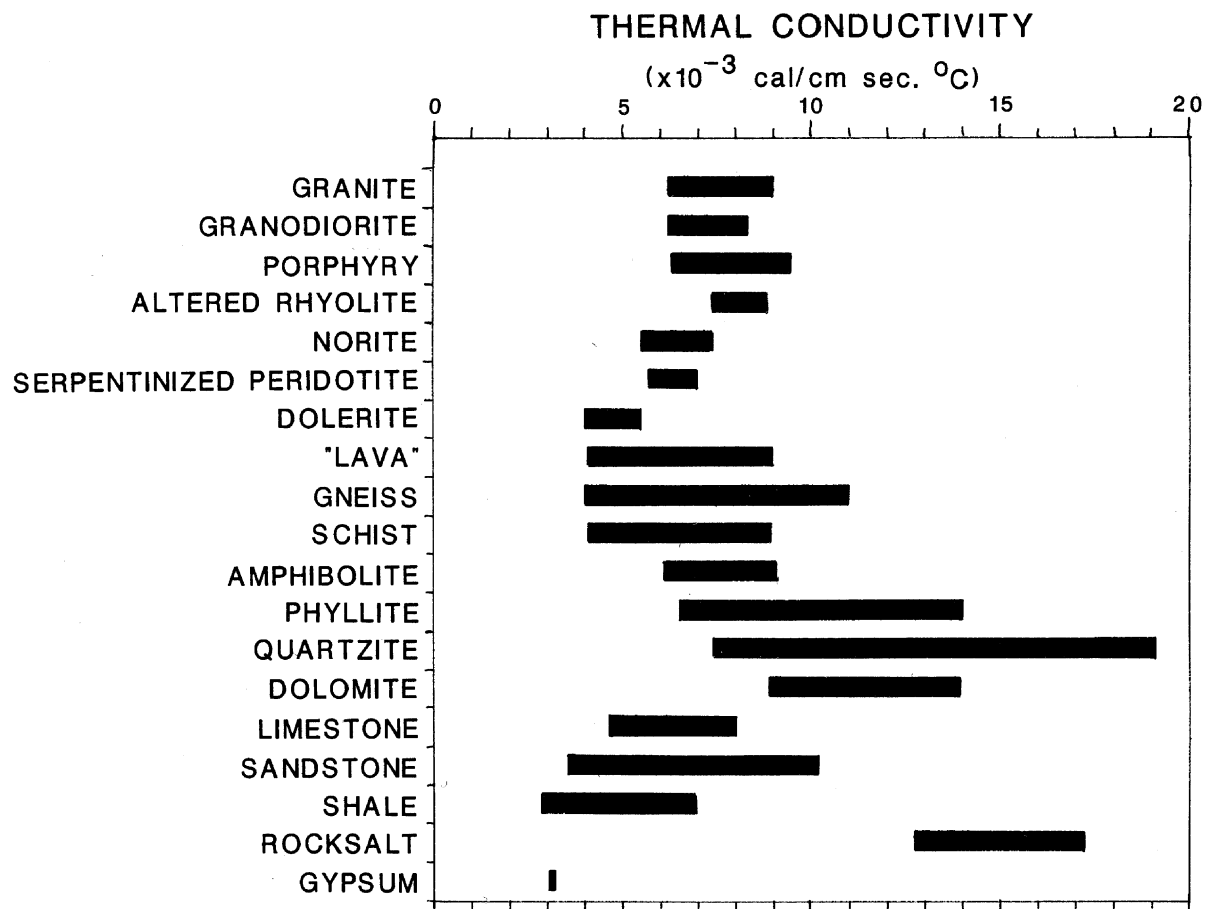


Figure 24. Thermal conductivities for selected sedimentary rocks from Clark (1966).

Many sedimentary and metamorphic rocks show distinct anisotropy, with thermal conductivities varying by as much as three times transverse or parallel to the grain (Heiland, 1940 and Clark, 1966).

The Debye relationship given by solid state theory provides a relationship among density, seismic velocity and thermal conductivity for crystalline solids (Kittel, 1956). This relationship is

$$K = 1/3 C v \Omega = 1/3 c \rho v \Omega \quad (26)$$

where C = specific heat per unit volume
v = seismic ρ velocity
 Ω = phonon mean free path
c = specific heat per unit mass
 ρ = density

Table 8. Thermal conductivities (K) for various sedimentary rocks as a function of water content, values in 10^{-3} calories/cm. sec. °c adapted from Clark (1966).

| Rock Type | Porosity | % moisture | K | % moisture | K |
|-------------|---|----------------------------|--|----------------------------|---|
| Limestone | 3.4 13.2 | air dry | 5.03 4.40 | saturated | 6.19 4.97 |
| Dolomite | 1.7 | " | 7.10 | " | 8.02 |
| Sandstone | 0.5 3.0 15.5 22. 29. 59. | " " " " " " | 9.20 15.5 7.26 4.43 3.69 1.26 | " " " " " " | 10.4 17.7 14.01 6.02 9.67 4.86 |
| Pumice | - | " | 0.60 | " | 1.2 |
| Sand | - | 0.2 | 0.65 | 30. | 3.94 |
| Loam | - | 0.3 | 0.79 | 27. | 5.50 |
| Clay | - | 1.4 | 0.57 | 67. | 3.6 |
| Frozen soil | - | 20. | 0.83 | 60. | 2.71 |

The Debye relationship has been used by Russian investigators (Neprimerov and others, 1989) to give estimates of thermal conductivities of crustal rocks based on their seismic velocities. Heiland (1940) gives a modified form for this relationship as follows:

$$K = Av^2\rho^2 \quad \text{where } A \text{ is a constant} \quad (27)$$

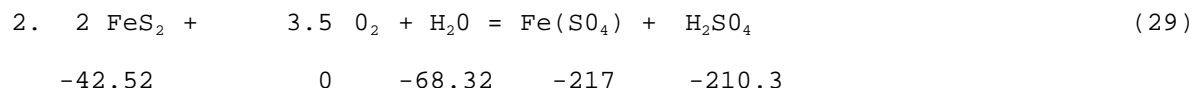
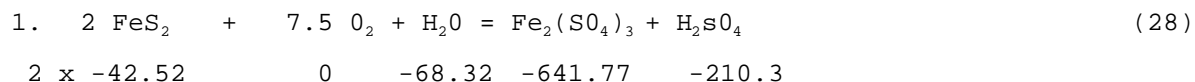
Heat sources

At present, radioelement decay constitutes the principal source of heat production for the earth as a whole, being approximately equal to that lost (Durrance, 1986). Quantitatively, the radioelement heat sources are much more significant than sources related to oxidizing sulfides. Radiogenic heat production from the Carnmenellis tin granite of SW England has been sufficient to account for all the heat necessary to drive the hydrothermal systems responsible for mineralization in the district (Durrance, 1986).

Only one locality where a natural fission reaction has occurred is known, that being the Oklo deposit in Gabon (Cowan, 1976). The estimated energy output of this natural reactor was 15,000 megawatt-years at a rate of 10 to 100 kilowatt per sec. Six individual reactor zones are known at the deposit. These zones are estimated to have gone critical about 1800 Ma ago, operating for about 0.5 Ma (Durrance, 1986).

Assuming standard isotopic ratios, the heat production from the three radioelements are, uranium 0.73, thorium 0.20, and potassium 2.7×10^{-5} calories per gram-year (Wetherill, 1966). Thus, for most rocks, which have a thorium to uranium ratio of 3.5:1 to 4:1, the heat contribution due to the thorium and uranium content is about equal. The contribution due to potassium for most rocks is typically several times smaller than that due to uranium or thorium. Table 9 gives values of heat production for various rock types whose radioelement composition is give by Vavilin and others (1982).

Locally, heat produced by oxidizing sulfides can be important. The quantity of heat generated is dependant on the particular sulfide, and the end products of oxidation. The rate of heat production is directly related to the rate of oxidation. For particular oxidation reactions, the heat generated can be calculated from values of heats of formation given in thermochemical compilations. The oxidation of pyrite to ferrous or ferric sulfate plus sulfuric acid provides an example to show the order of magnitude of heat generated. The reactions are:



Heats of formation in kilo-calories per mole are indicated below each reactant or product, with the products assumed in dilute solution. For reaction 1 the heat produced is 699 kcal or 2.91 kcal/gm of pyrite, and for reaction 2, 316 kcal or 2.64 kcal/gm of pyrite.

Table 9. Average radiogenic heat production for various types of rocks.
Compositional data from Vavilin and others (1982).

| Lithology | Composition | | | Heat Production cal/gm 10^6 yr. | | | |
|------------------------------------|-------------|-------|--------|-----------------------------------|------|------|-------|
| | K% | u ppm | Th ppm | K | U | Th | Total |
| Peridotite pyroxenite | 0.012 | 0.03 | 0.08 | 0.0032 | .022 | .016 | 0.041 |
| Gabbro- diabase | 0.56 | 0.6 | 1.8 | 0.15 | 1.44 | .36 | 0.95 |
| Quartz diorite- granodiorite | 2.05 | 2*1 | 8.3 | 0.55 | 1.53 | 1.66 | 3.74 |
| Plagiogranite | 1.26 | 2.7 | 9.6 | 0.34 | 1.97 | 1.92 | 4.23 |
| Granite | 4.43 | 4.5 | 18.0 | 1.20 | 3.28 | 3.60 | 8.08 |
| Alkali- granite | 3.90 | 6 | 25. | 1.05 | 4.38 | 5.00 | 10*43 |
| | | | | | | | |
| basalt- diabase | 0.55 | 0.7 | 2.3 | 0.15 | 0.51 | 0.46 | 1.12 |
| andesite | 1.66 | 1.2 | 4.0 | 0.45 | 0.88 | 0.80 | 2.13 |
| dacite | 2.20 | 2.5 | 10.0 | 0.59 | 1.82 | 2.00 | 4.41 |
| rhyolite | 3.87 | 4.7 | 19.0 | 1.04 | 3.43 | 3.80 | 8.27 |
| | | | | | | | |
| gravel- conglomerate | | 2.4 | 9.0 | - | 1.75 | 1.80 | 3.55 |
| sandstone- siltstone | 1.07 | 2.9 | 10.4 | 0.29 | 2.12 | 2.08 | 4.49 |
| clay- argillite | 2.66 | 4.0 | 11.5 | 0.72 | 2.92 | 2.30 | 5.94 |
| limestone | 0.27 | 1.6 | 1.8 | 0.073 | 1.17 | 0.36 | 1.60 |
| dolomite | 0.27 | 3.7 | 2.8 | 0.073 | 2.70 | 0.56 | 3.33 |
| chert- quartzite | trace | 1.7 | 2.2 | trace | 1.24 | 0.44 | 1.68 |
| evaporates | trace | 1.0 | 1.0 | trace | 0.73 | 0.20 | 0.93 |
| | | | | | | | |

RADIOELEMENT CONTENT

The radioelements K, U, and Th in rocks are present either as essential elements in a variety of minerals present in a rock, or substituting, often as trace elements, in minerals. Potassium, for most rocks, is present in the potash feldspars, microcline and orthoclase or in micas such as muscovite and biotite. In evaporates, the minerals sylvite and carnallite can be important sources of potassium. In epithermal deposits, hydrothermal alteration assemblages may include alunite, sericite, and adularia as the principal potassium-containing species of mineral.

Uranium and thorium in igneous and metamorphic rocks are typically present in the accessory minerals apatite, sphene, and zircon, and in the rarer species allanite, monazite, pyrochlore, thorite, uraninite and xenotime. In the near surface oxidizing environment, uranium tends to be very mobile, dissolving in the ground water and then being deposited where conditions are more reducing. Carnotite, uranophane, and other secondary uranium minerals are of this type. Thorium, on the other hand, is relatively stable, remaining with the resistate minerals.

Compilations of the Radioelement contents of rocks are rather sparse in the literature. Data from Vavilin and others (1982) were given in table 9 where radiogenic heat production was discussed. The compositional data given there can be compared with table 10 which shows some values obtained on various U.S. Geological Survey standard rocks and several average rock types. A prominent feature seen in table 10 is the abrupt decrease in Radioelement content between mafic and ultramafic rocks.

Another feature evident in table 10 is the general correspondence between potassium and thorium. Portnov (1987) discusses this relationship in detail. As the ratio K(%) / Th(ppm) is typically 0.17 to 0.20, he calls rocks having ratios significantly different from this range either thorium or potassium specialized. Igneous rocks or derived metasomatites showing potassium specialization are identified with gold-silver, silver-polymetallic, molybdenum, and bismuth deposits. Thorium specialized rocks and metasomatites

Table 10. Radioelements content of selected USGS rock standards and of other rocks. Data from Van Schmus (1984).

| Rock Type | Composition | | |
|-------------------------|-------------|-------|--------|
| | K% | U ppm | Th ppm |
| GSP-1 granodiorite | 4.50 | 2.0 | 104 |
| G-1 granite | 4.45 | 3.4 | 50 |
| Ave. Low-Ca granite | 4.20 | 4.7 | 20 |
| G-2 granite | 3.67 | 2.0 | 24 |
| STM-1 nepheline syenite | 3*54 | 9.1 | 27 |
| RGM-1 rhyolite | 3.49 | 5.8 | 13 |
| MAG-1 marine mud | 2.96 | 2.8 | 12.2 |
| QLO-1 quartz latite | 2.90 | 5.8 | 13 |
| SDC-1 mica schist | 2.71 | 3.1 | 11.4 |

Table 10. Radioelements content of selected USGS rock standards and of other rocks. Data from Van Schmus (1984).

| Rock Type | Composition | | |
|------------------|-------------|-------|--------|
| | K% | U ppm | Th ppm |
| Ave. shale | 2.66 | 3.7 | 12. |
| AGV-1 andesite | 2.35 | 1.9 | 6.4 |
| SCO-1 shale | 2.20 | 3.1 | 9.5 |
| BCR-1 basalt | 1.38 | 1.7 | 6.0 |
| Ave. sandstone | 1.07 | 1.7 | 5.5 |
| Ave. basalt | 0.83 | 0.9 | 2.7 |
| W-1 | 0.52 | 0.6 | 2.4 |
| BHVO-1 basalt | 0.43 | 0.5 | 0.9 |
| Ave. carbonate | 0.27 | 2.2 | 1.7 |
| Ave. ultramafic | 0.003 | 0.001 | 0.004 |
| DTS-1 dunite | 0.001 | 0.004 | 0.01 |
| PCC-1 peridotite | 0.001 | 0.005 | 0.01 |

are identified with tin, tungsten, rare-earth, and rare-metal deposits. These characteristics thus can be used to assess the favorability of regions for one or the other group of mineral deposits.

Hansen (1980) and Wollenberg and Smith (1987) have presented average values and ranges of Radioelement content for a variety of rocks. These data are shown in figure 25. These data indicate the potential of gamma-ray spectrometry in lithologic discrimination where element concentration and element ratios may be combined to give a powerful mapping tool.

Soils may be depleted in Radioelement content, particularly in uranium and potassium, dependent on the weathering environment, and host mineralogy of the radioelements. Soils in tropical regions are particularly prone to leaching of most soluble components. However, as Guillemot (1988) has shown good quality radioelement data, capable of mapping lithologies, can be obtained even in tropical areas.

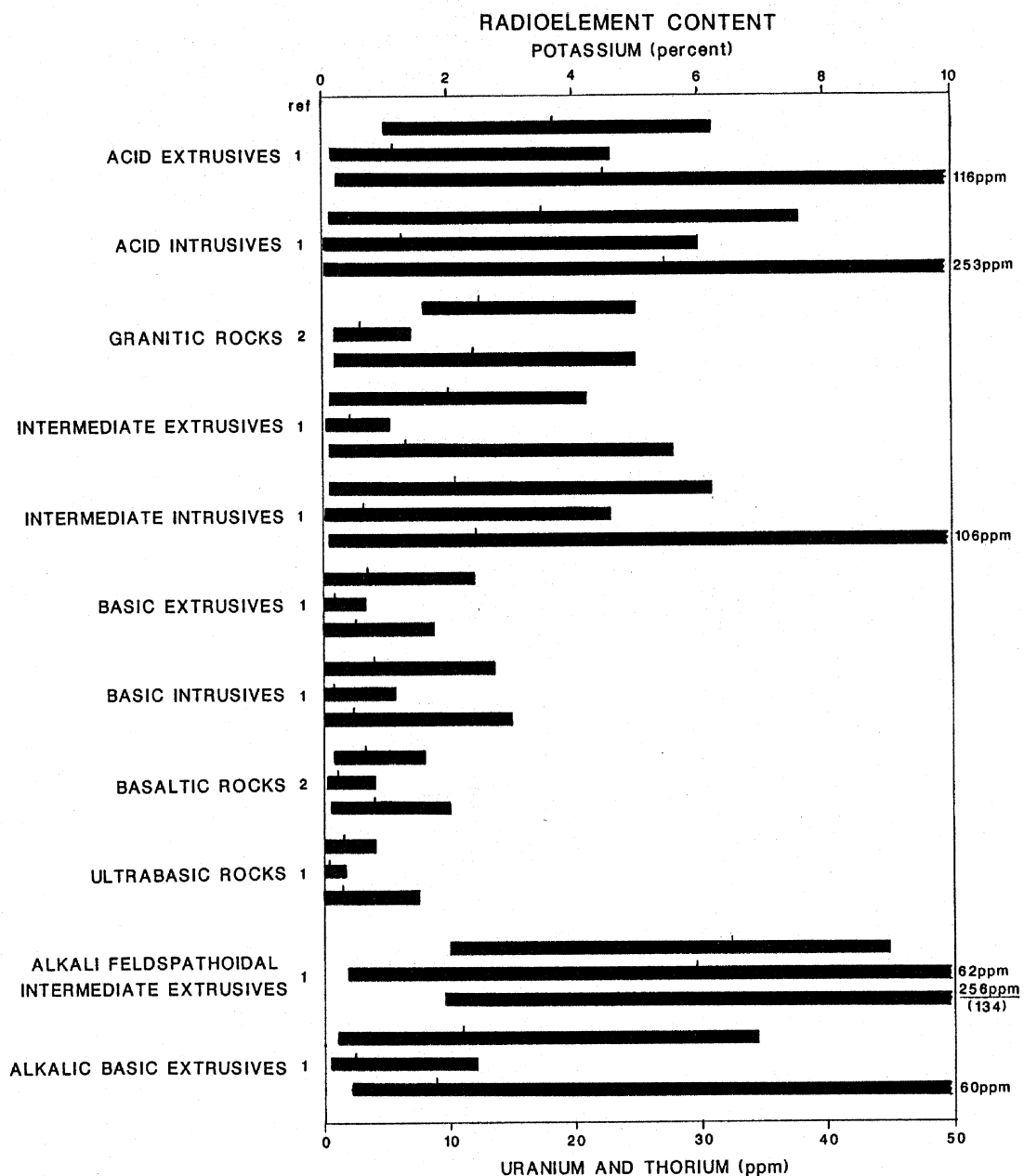


Figure 25. Radioelement contents reported for a variety of lithologies from 1. Wallenberg and Smith (1982 and 2. Hansen (1980). For each type of lithology the elements are in the order top to bottom, K, U, Th. The small vertical bar indicates the mean value.

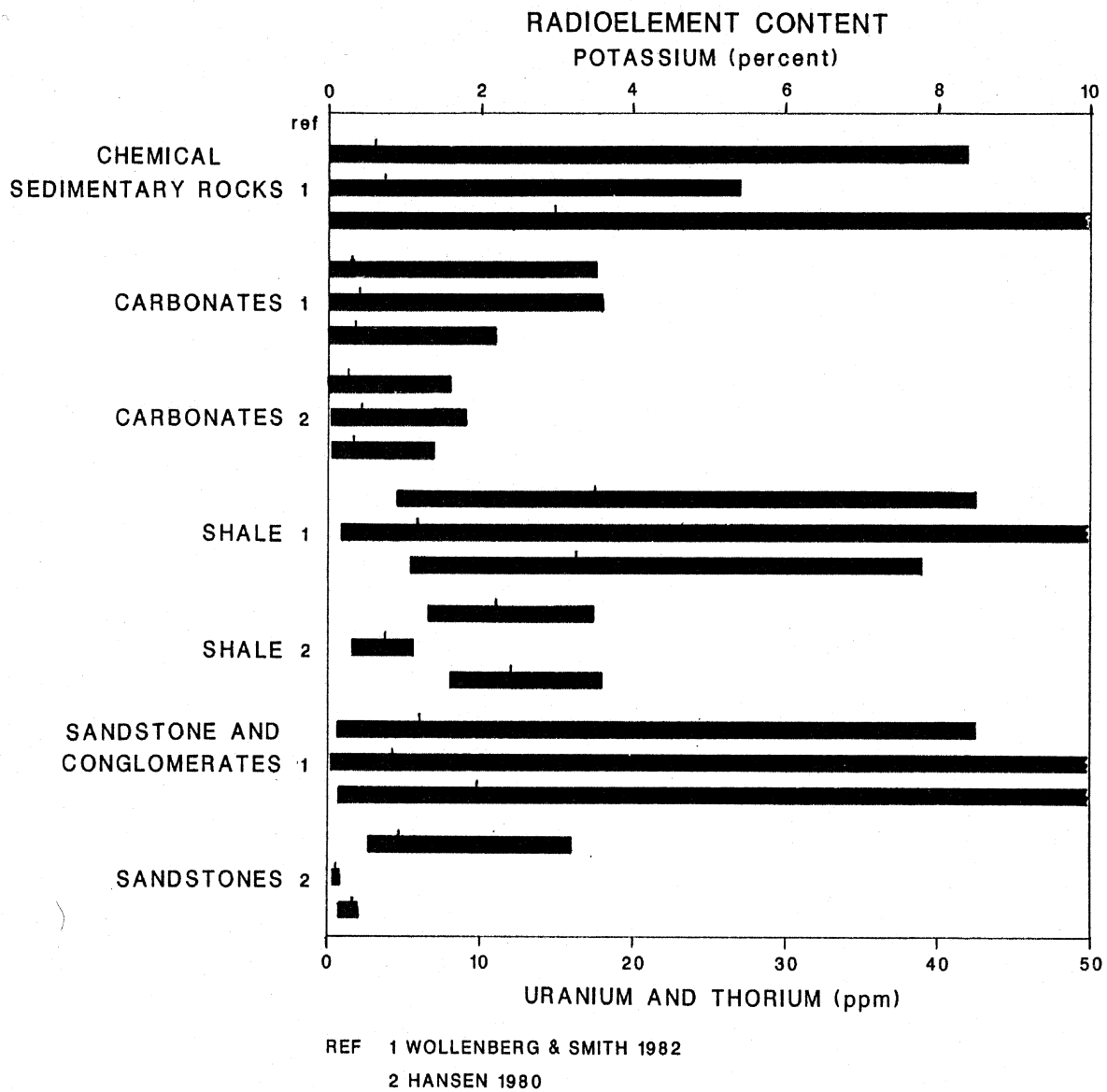


Figure 25 (continued)